# The variety and distribution of submarine glacial landforms and implications for ice-sheet reconstruction


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Glacimarine processes affect about 20% of the global ocean today, and this area expanded considerably under cyclical full-glacial conditions during the Quaternary (Fig. 1) (Dowdeswell *et al.* 2016*b*). Many of the submarine landforms produced at the base and margin of past ice sheets remain well preserved on the seafloor in fjords and on high-latitude continental shelves after the retreat of the ice that produced them. These glacial landforms, protected from subaerial erosion and beneath wave-base and tidal currents in water that is often hundreds of metres deep, are gradually buried by both hemipelagic and glacimarine sedimentation; they may be preserved over long periods in the geological record if palaeo-continental shelves are not reworked by subsequent glacier advances or bottom currents (Dowdeswell *et al.* 2007). This means that, first, submarine glacial landforms can be observed at or close to the modern seafloor after retreat of the last great ice sheets from their most recent Quaternary maximum about 18–20 000 years ago using swath-bathymetric mapping systems and, secondly, buried glacial landforms may also be identified and examined within glacial-sedimentary sequences from Quaternary and earlier ice ages using seismic-reflection methods.

The development of multibeam echo sounding over the past two decades, coupled with high-accuracy GPS positioning, has allowed morphological mapping of the seafloor at an unprecedented level of detail. In this paper, the variety of submarine glacial landforms observed in modern, Quaternary and more ancient sediments is described. Landforms produced subglacially, those formed at and beyond marine ice-sheet margins, together with the slope processes and bottom currents likely to result in their reworking, are considered along with the sediment volumes of these landforms and the time they take to develop. This is followed by discussion of the significance of submarine glacial landforms and landform-assemblages for reconstruction of the form and flow of past ice sheets, including the former extent and flow direction of ice sheets, whether they are flowing fast as ice streams or slowly in inter-ice stream areas, the nature and rate of ice-sheet deglaciation, conditions at past ice-sheet beds, and inferences on the characteristics of the basal hydrological system and past climatic conditions.

## The variety of submarine glacial landforms

### *Subglacial landforms*

Landforms produced at the ice–bed interface by subglacial processes (Table 1) are usually streamlined and elongate in shape, with their long axes orientated in the direction of ice flow; this is typical for both sedimentary landforms linked to soft-sediment deformation (e.g. Clark 1993; Canals *et al.* 2000; Ó Cofaigh *et al.* 2002; Spagnolo *et al.* 2014) and those shaped by the erosion of bedrock (e.g. Dahl 1965; Krabbendam *et al.* 2016; Nitsche *et al.* 2016*a*). In addition, some glacial landforms have distinctive shapes differentiating their up-flow (or stoss) and down-flow (lee) sides, with the up-flow end usually being steeper and/or less elongate than a down-flow tail that often tapers in the direction of flow; relatively blunt-nosed drumlins are a clear example (e.g. Clark *et al.* 2009; Spagnolo *et al.* 2010).

There are many morphological variations among the streamlined and elongate sedimentary landforms produced at the base of ice sheets (e.g. Dunlop & Clark 2006; Benn & Evans 2010), particularly beneath fast-flowing ice streams (e.g. Canals *et al.* 2000; Spagnolo *et al.* 2014), and a number of theories exist concerning the detailed mechanisms by which they form (e.g. Hindmarsh 1999; Clark *et al.* 2003; Schoof & Clarke 2008; Spagnolo *et al.* 2016). These elongate subglacial landforms are often distinguished on the basis of their absolute size and elongation ratio (the ratio between their maximum length and width). Thus, elongate flutes are shorter than drumlins, which are in turn exceeded in length by mega-scale glacial lineations (MSGLs). In addition, Rogen moraines may represent an early stage in the streamlining of what may originally have been transverse-to-flow ridges of sediment (Dunlop & Clark 2006).

Individual streamlined landforms are often found in clusters or fields of similar morphology, where individual landforms are distributed with their long axes parallel or sub-parallel to one another. In addition, multiple generations of such sedimentary landforms, of differing orientation, may be superimposed on one another, recording shifts in the direction of past ice flow (e.g. Clark 1993; Greenwood *et al.* 2012). It is also important to point out that although most morphological analyses of streamlined subglacial landforms have taken place using Quaternary examples from terrestrial and marine settings, seismic-reflection and ice-penetrating radar data demonstrate that streamlined sedimentary landforms are being produced today beneath actively flowing ice streams in Antarctica (e.g. King *et al.* 2007, 2009; Smith & Murray 2009).

In the glacimarine environment, streamlined sedimentary and bedrock landforms have been observed at or close to the seafloor on many formerly ice-covered continental shelves. Swath-bathymetric images of submarine Rogen moraines, drumlins and MSGLs provide examples of streamlined and increasingly elongate sedimentary landforms (Fig. 2a–c), whereas streamlined bedrock landforms are made up in whole or in part of indurated rock and sculpted over probably much longer periods by subglacial erosion processes (Fig. 2d). The way that fields of blunt-nosed drumlins and more elongate MSGLs cluster and are orientated consistently parallel or sub-parallel to one another can also be seen (Fig. 2b, c). Some extreme examples of subglacial streamlining have been reported from Barents Sea sediments where large needle-like features, of much greater volume than classical MSGLs (Spagnolo *et al.* 2014), have elongation ratios of up to 65:1 (e.g. Bjarnadóttir *et al.* 2014; Bjarnadóttir & Andreassen 2016*a*).

Glacial landforms interpreted to have formed at the lateral margins of former ice streams, where shear margins represent a well-defined boundary with slower-flowing ice to either side (e.g. Bentley 1987), include lateral shear-margin moraines that can reach tens of kilometres in length and tens of metres in height


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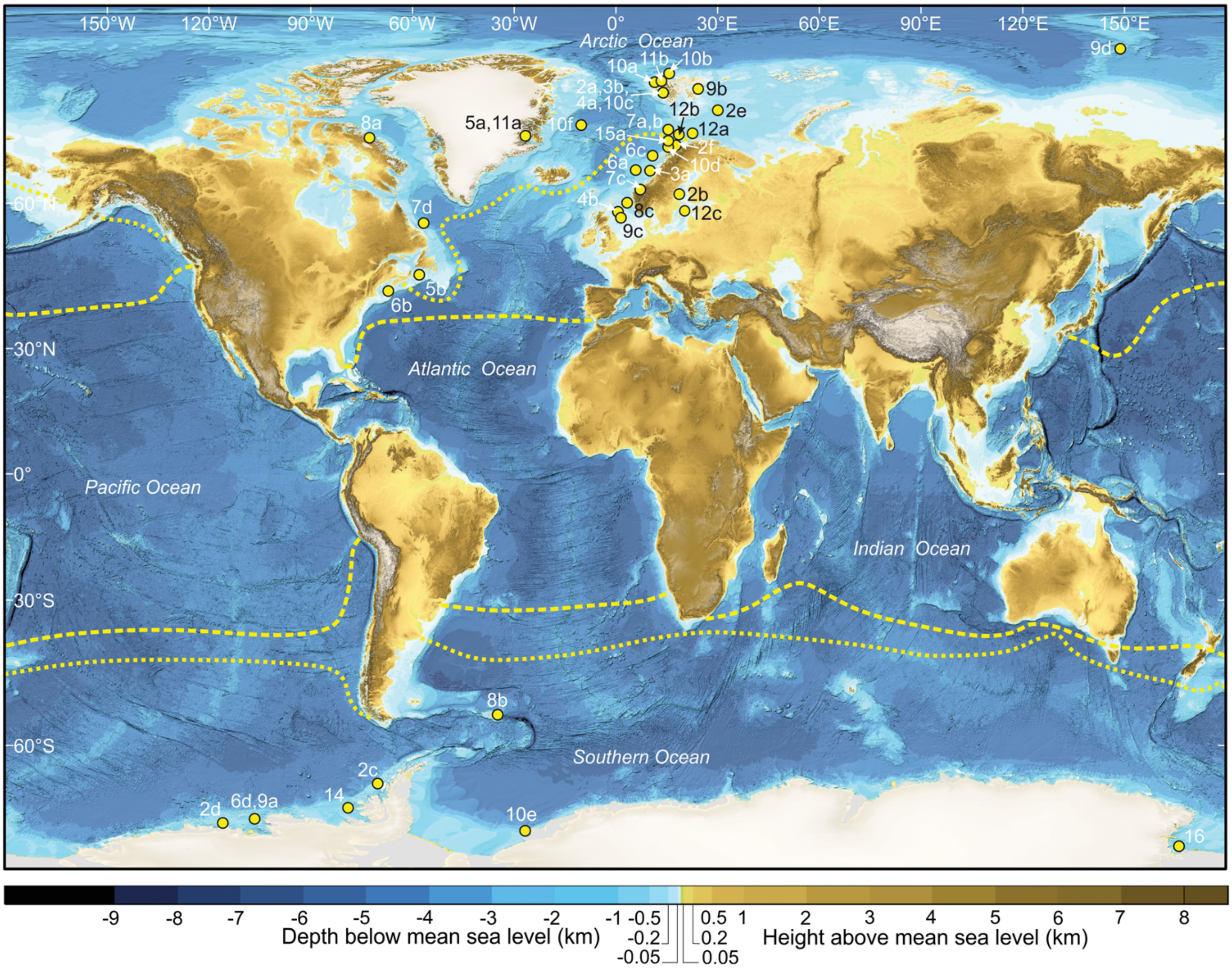


**Fig. 1.** The global distribution of glaciers and ice sheets and the glacier-influenced, or glacimarine, environment. The approximate modern (yellow dotted line) and Quaternary full-glacial (yellow dashed line) limits of ice-rafting and ice-keel ploughing of the seafloor are shown (modified from Anderson 1983). GEBCO World Map: Gall projection. Numbered yellow dots refer to the locations of subsequent figures.

(e.g. Stokes & Clark 2002*a*; Batchelor & Dowdeswell 2016). Strings of apparently rafted glacitectonic blocks, observed in Barents Sea sediments and orientated in the direction of past ice flow, are also thought to be related to the lateral margins of past ice streams where oblique-to-flow shear takes place under compressional strain (Rüther *et al.* 2013, 2016). Aligned strings of glacitectonic rafts and lateral shear-margin moraines are shown in Figure 2e and f.

Not all subglacially produced landforms are streamlined and elongate, however (Fig. 3; Table 1). An example is the so-called hill–hole pair (Fig. 3a). The depressions forming the ‘holes’ occur up-flow of sedimentary ‘hills’ of approximately equivalent dimensions, with ice-sheet basal processes enabling sediment transport between the hole and hill (*c.* 10–30 m for both height and depth). Typical hill–hole-pair volumes are up to about a cubic kilometre (e.g. Rise *et al.* 2016). Several hill–hole pairs have been described from the Norwegian shelf (Sættem 1990; Ottesen *et al.* 2005*a*), and similar features located on land in the Canadian Prairies are made up of thrust blocks probably formed close to former glacier termini (e.g. Moran *et al.* 1980). The compressive stresses thought to be needed to produce tectonism are probably related to basal freezing close to relatively thin ice margins. Some hill–hole pairs have streamlined sedimentary tails, implying reworking by subsequent ice reactivation or readvance.

In addition, sets of complex subglacially formed sedimentary ridges, a few metres in height and spaced a few hundred metres apart, provide a reticulate or box-work pattern on the seafloor and are sometimes referred to as rhombohedral moraines (Fig. 3b) (Solheim & Pfirman 1985). These distinctive sets of submarine glacial landforms are interpreted to have formed at the base of largely stagnant ice, as soft subglacial sediment is squeezed up into basal crevasses. This process is inferred to take place particularly at the termination of the active, fast-flowing stage of the surge cycle in surge-type glaciers (e.g. Meier & Post 1969; Kamb *et al.* 1985; Cuffey & Paterson 2010). At this time, rapid reorganization of the basal hydrological system enables efficient meltwater drainage, leading to low basal water pressure and the squeezing of soft sediment into crevasses formed during active surging of the overlying ice (Solheim & Pfirman 1985). Such sets of submarine glacial landforms have been observed on the seafloor close to the retreating termini of a number of surge-type glaciers, especially the tidewater glaciers of Svalbard (Solheim & Pfirman 1985; Ottesen & Dowdeswell 2006; Ottesen *et al.* 2008; Flink *et al.* 2015). Readvances during subsequent active phases of the surge cycle may rework and elongate the ends of these ridges into forms that

**Table 1.** *The variety of submarine landforms produced in glacial and glacimarine environments*

| Process environment | Glacial or glacimarine landform |
|---|---|
| *Subglacial* | Mega-scale glacial lineations |
| | Flutes and drumlins |
| | Crag-and-tails |
| | Ice-moulded bedrock |
| | Hill–hole pairs |
| | Crevasse-fill ridges |
| | Tunnel valleys (glacifluvial) |
| | Bedrock channels (glacifluvial) |
| | Eskers (glacifluvial) |
| | Fjords |
| | Cross-shelf troughs |
| *Ice-marginal* | Terminal and recessional moraine ridges |
| | Hummocky-terrain belts |
| | Small retreat moraines |
| | Grounding-zone wedges |
| | Ice-proximal fans |
| | Ice-stream lateral moraines |
| | Trough-mouth fans |
| *Glacimarine* | Iceberg-keel ploughmarks |
| | Sea-ice keel ploughmarks |
| | Smooth basin-fill from meltwater plumes |
| *Marine* | Wave and current features (inc. turbidity currents) |
| | Mass movements (inc. slides and debris flows) |
| | Fluid-escape features |
| | Features related to human activity |

resemble Rogen moraines mapped on land (e.g. Dunlop & Clark 2006; Ottesen & Dowdeswell 2006).

Subglacial landforms produced by meltwater flow are a further exception to the notion that subglacial landforms are necessarily streamlined (Fig. 4; Table 1). Subglacial channels and their sedimentary fill, including both eskers and the very large tunnel valleys found in Quaternary and ancient glacial rocks (e.g. Boyd *et al.* 1988; Huuse & Lykke-Andersen 2000; Hirst *et al.* 2002; Le Heron *et al.* 2004; Stewart & Lonergan 2011), tend to be sinuous and often branching or anabranching in planform; mean subglacial channel orientation may, nonetheless, be in the general direction of ice flow given the control on subglacial water-pressure gradient exerted by ice-surface slope in particular (e.g. Shreve 1972, 1985). Such channels, which are preserved in both sedimentary and rock beds, and may also have long profiles that include both positive and negative slopes, are formed by water flowing under pressure beneath past ice sheets (e.g. Ó Cofaigh *et al.* 2002; Lowe & Anderson 2003; Domack *et al.* 2006; Smith *et al.* 2009).

At broader spatial scales, fjords cut into bedrock and cross-shelf troughs usually eroded into sediments represent two types of very large subglacially produced landform (Fig. 5; Table 1). Fjords of characteristic U-shape in cross-section may be tens to hundreds of kilometres long, sometimes over 1000 m deep and occur on all mountainous high-latitude margins (Fig. 5a) (e.g. Holtedahl 1967; Syvitski *et al.* 1987; Barnes *et al.* 2016; Dowdeswell *et al.* 2016*a*). They have developed as glacier-ice eroded pre-glacial river systems over successive Quaternary full-glacial periods, and many also have a substantial structural–geological influence (Lind & Andrews 1985; Glasser & Ghiglione 2009). The rate of development of fjords by subglacial erosion has been modelled by Harbor (1992).

Curvilinear cross-shelf troughs are up to several hundred metres deep relative to the surrounding continental shelf, tens of kilometres wide and tens to hundreds of kilometres long; such cross-shelf troughs are very common morphological features of Arctic and Antarctic margins (e.g. Canals *et al.* 2002; Ottesen *et al.* 2005*a*; Livingstone *et al.* 2012; Batchelor & Dowdeswell 2014). Well-known examples include the series of cross-shelf troughs on the Norwegian margin, including the Norwegian Channel and the Bear Island Trough (Ottesen *et al.* 2005*a*; Andreassen *et al.* 2014), and the Laurentian Channel between Nova Scotia and Newfoundland (Fig. 5b) (Todd 2016*a*). These major submarine landforms are interpreted to have been produced by subglacial erosion of underlying sediments at the base of fast-flowing ice streams. The increasing availability and examination of seismic-reflection data, in some cases in both 2D and 3D form, from the Norwegian margin in particular, have demonstrated that ice-stream positions may change between and within glacial cycles of the Quaternary, implying significant shifts in the focus of erosion and sediment delivery to the shelf and shelf edge (e.g. Dowdeswell *et al.* 2006; Sarkar *et al.* 2011; Winsborrow *et al.* 2012). It is also known from modern satellite observations of West Antarctica that the ice streams draining into the Ross Sea region have undergone dynamic changes and even switch-off over the last few decades to centuries (e.g. Conway *et al.* 2002; Joughin *et al.* 2002).

### *Ice-marginal landforms: line-sourced ridges*

Submarine landforms produced at either grounded tidewater glacier margins or at the grounding zone of floating ice shelves are typically orientated transverse to the direction of ice flow and are provided with sediment from a line source that is the length of the marine-terminating ice mass (Fig. 6; Table 1); this is a clear contrast with the parallel-to-flow orientation of many subglacial landforms (e.g. Ottesen & Dowdeswell 2009).

Terminal, recessional and smaller moraine ridges, sometimes produced annually, each form along the grounded marine margins of glaciers and ice sheets. The sizes of these moraine ridges, which are often asymmetrical with steeper ice-distal faces (Benn & Evans 2010), are dependent on the rate of sediment supply from the parent glacier and the time that the ice margin has been stationary at a given location. There may be a tectonic push component to some ridges if minor changes in ice-margin location take place. Terminal moraine ridges, marking the outermost limit of ice-sheet margins during full-glacial conditions at the shelf edge, can reach over 100 m above the surrounding seafloor and tens of kilometres in length; the huge Skjoldryggen Ridge west of Norway is a particularly well-developed example (Fig. 6a) (Andersen 1979; Dowdeswell *et al.* 2016*g*).

Recessional moraines are similarly line-sourced, but are produced during stillstands in the regional deglacial retreat of grounded tidewater glacier margins that do not have a floating ice shelf beyond the grounding zone. The presence of vertical ice cliffs, rather than the narrow basal cavities associated with the grounding zone of floating ice shelves (e.g. Dowdeswell & Fugelli 2012; Horgan *et al.* 2013), provides the accommodation space needed to build ridges that may be tens of metres high. Significant pauses in regional deglaciation are often associated with topographic highs or lateral constrictions in cross-shelf troughs and fjords; recessional ridges are built at such locations. Topographic shallows reduce water depth and buoyancy at the ice margin and lateral constrictions provide additional back-stress, both of which act to slow the rate of iceberg production and, hence, the rate of retreat of marine-terminating ice sheets, causing halts in ice recession (e.g. Jamieson *et al.* 2012).

Sets of smaller transverse ridges on the seafloor, often only a few metres high and spaced 50 m to a few hundred metres apart (Fig. 6b), are common in relatively shallow shelf and fjord settings where deglaciation takes the form of the slow retreat of a grounded tidewater glacier margin (e.g. Boulton 1986; Ottesen & Dowdeswell 2006; Dowdeswell *et al.* 2008). Individual small ridges form at the ice margin regularly, often annually, as the ice front retreats, and some ridges are thought to be produced during minor winter readvances of the ice front, induced by a lack of iceberg calving in the presence of sea ice, superimposed on more general summer retreat. Sometimes groups of tens to hundreds of small

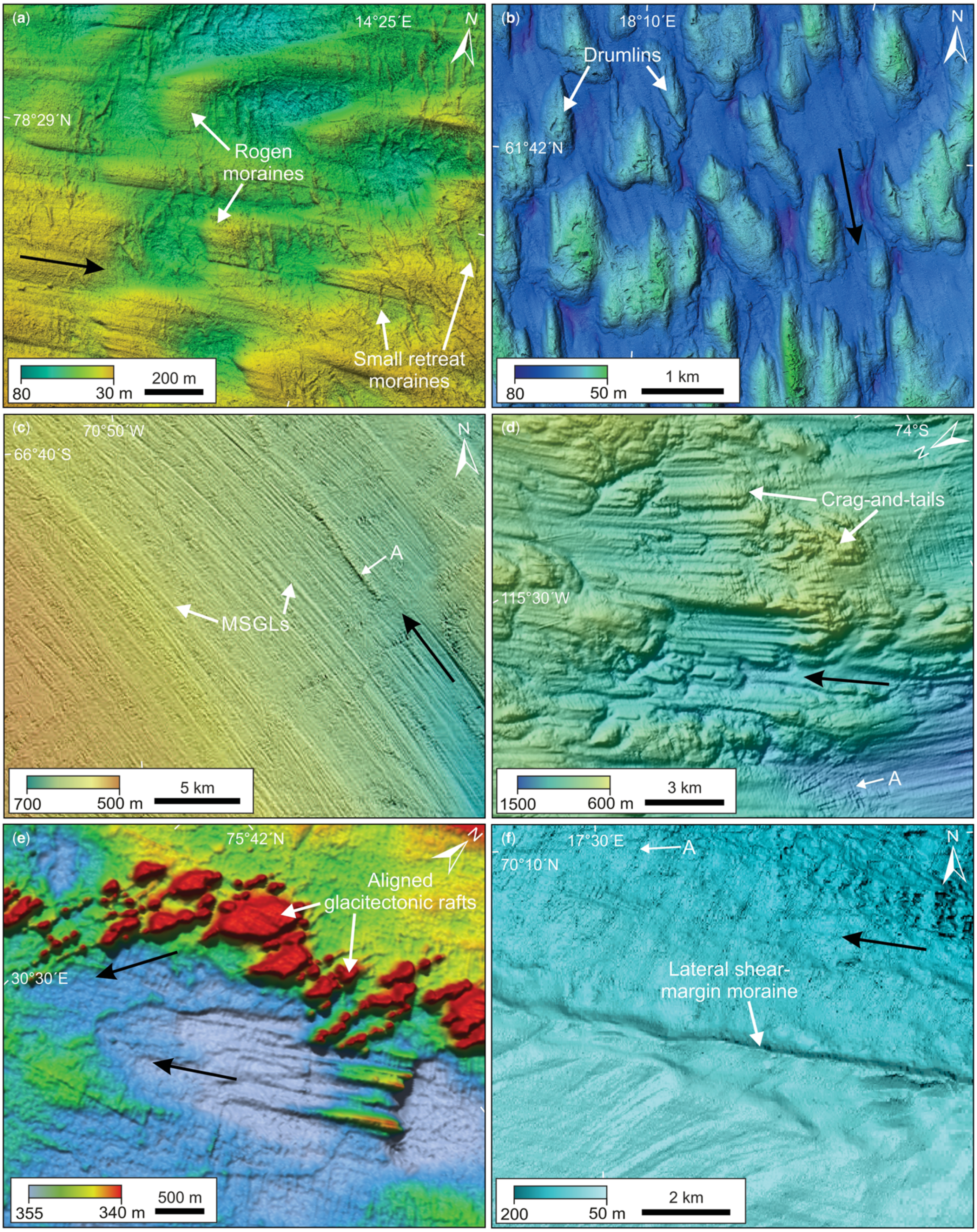


**Fig. 2.** Submarine examples of streamlined subglacial landforms and those located at the lateral margins of ice streams. (**a**) Streamlined Rogen moraines with superimposed transverse-to-flow small retreat ridges in a Svalbard fjord (modified from Dowdeswell & Ottesen 2016*a*). (**b**) Part of a drumlin field in the Gulf of Bothnia (modified from Jakobsson *et al.* 2016*a*). (**c**) MSGLs on the floor of Marguerite Trough, Antarctic Peninsula (modified from Ó Cofaigh *et al.* 2016*a*). (**d**) Crag-and-tails on the Amundsen Sea shelf, Antarctica (modified from Nitsche *et al.* 2016*a*). (**e**) String of aligned glacitectonic rafts in the Barents Sea (modified from Rüther *et al.* 2016). (**f**) Lateral shear-margin moraine at the edge of a Norwegian cross-shelf trough (modified from Batchelor & Dowdeswell 2016). Several artefacts are labelled A. Black arrows show the direction of past ice flow.

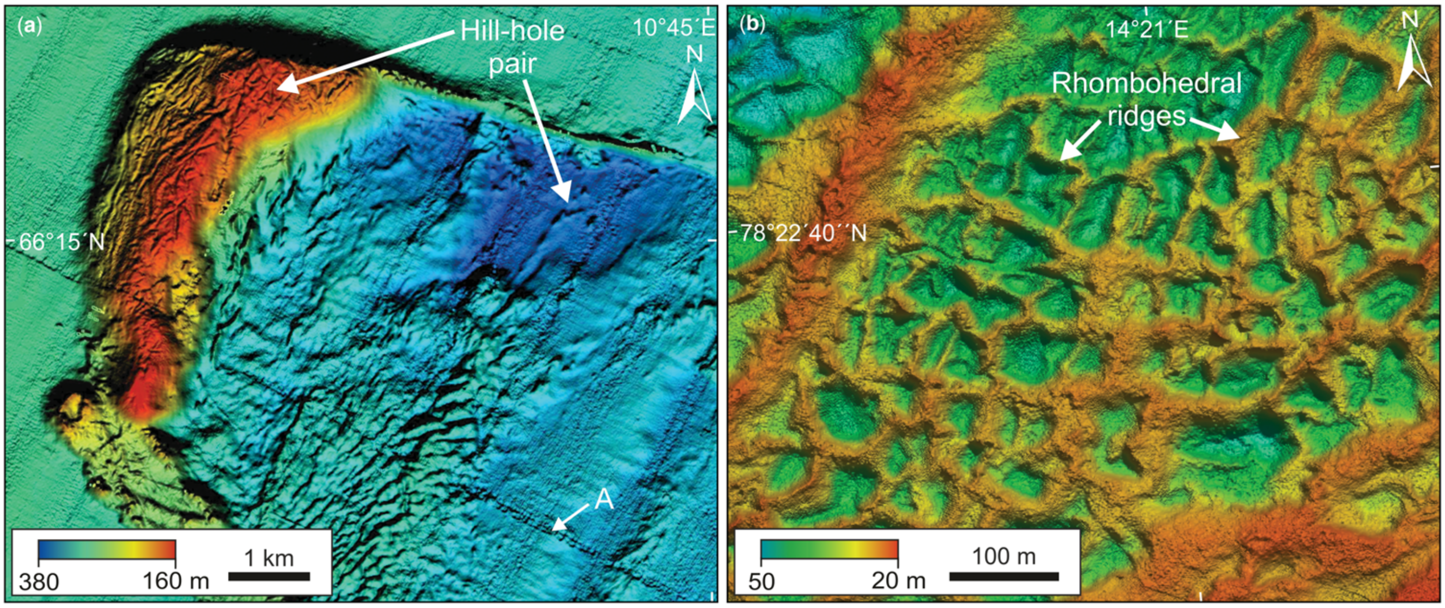


**Fig. 3.** Submarine examples of non-streamlined subglacial landforms. (**a**) Hill–hole pair on Trænabanken, west of Norway (modified from Dowdeswell *et al.* 2016*h*). A is an example of an artefact. (**b**) Rhombohedral pattern of crevasse-fill ridges in a Svalbard fjord (modified from Dowdeswell & Ottesen 2016*a*).

retreat moraines are found at the seafloor (e.g. Shipp *et al.* 1999; Dowdeswell *et al.* 2008). In several Spitsbergen fjords, comparisons between the known locations of past glacier termini over the past century or so and submarine moraine ridges have demonstrated clearly that the ridges are formed annually (e.g. Ottesen & Dowdeswell 2006; Burton *et al.* 2016), although this may not be so in all glacimarine settings. Similar ridges in marine and lacustrine environments, first observed in Scandinavia, have been referred to as De Geer moraines (e.g. De Geer 1889; Lindén & Möller 2005; Todd 2016*b*).

The interpretation of belts of hummocky seafloor topography at the shelf edge (Fig. 6c), associated with the maximum extent of the Eurasian Ice Sheet at the Last Glacial Maximum (LGM), is less clear-cut. The hummocky belts of sedimentary debris are a kilometre or so wide and extend for many kilometres along the shelf edge in several locations west of Norway and NW Svalbard (Ottesen & Dowdeswell 2009; Elvenes & Dowdeswell 2016). These landform belts, whose hummocks are of only a few metres to about 20 m in amplitude, are tentatively interpreted to be produced by the oscillating motion of an ice margin that is close to fully buoyant and moves up and down with the tide. In support of this suggestion, it has been demonstrated that the modern margins of ice sheets exhibit regular tidally generated vertical motion for kilometres to tens of kilometres inland of their grounding zones (e.g. Bindschadler *et al.* 2003; Gudmundsson 2006).

In high-latitude settings where grounded ice is fringed by a floating ice shelf attached to and fed by the parent ice sheet, the grounding zone may be a few kilometres to several hundred kilometres from the floating ice-shelf edge where icebergs are released (e.g. Dowdeswell & Bamber 2007). There are, therefore, large cavities beneath ice shelves that are filled with marine water often following well-defined ocean-circulation patterns

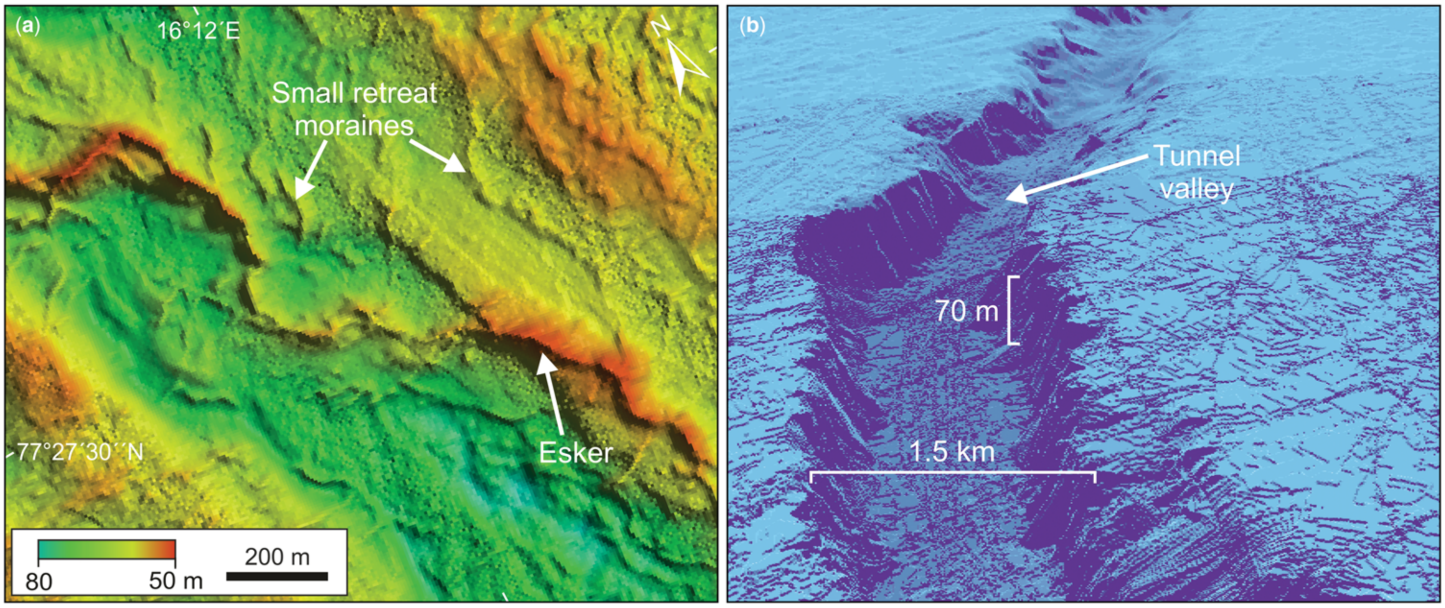


**Fig. 4.** Submarine examples of landforms produced by subglacial meltwater. (**a**) An esker, with small transverse ridges superimposed on it, in Svalbard fjord (modified from Dowdeswell & Ottesen 2016*a*). (**b**) Tunnel valley in the North Sea from Olex data (modified from Stewart 2016).

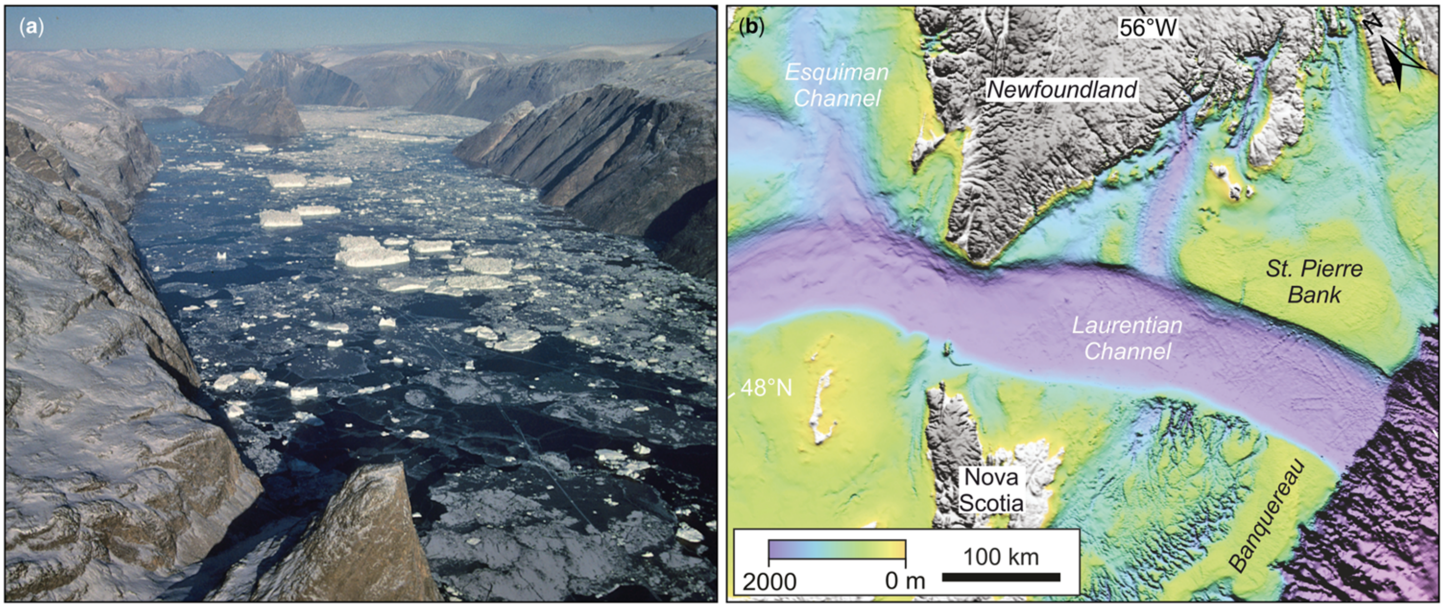


**Fig. 5.** Large subglacially eroded landforms. (**a**) The 5 km-wide and steep-walled Nordvestfjorden, inner Scoresby Sund, East Greenland (photo: J.A. Dowdeswell). (**b**) The Laurentian Channel: a cross-shelf trough on the Scotian Shelf, Atlantic Canada (modified from Todd 2016*a*).

(e.g. Mayer *et al.* 2000; Jenkins *et al.* 2010). Both theory and geophysical observations suggest that the part of the cavity closest to the grounding zone itself is narrow (e.g. Bindschadler 1993; Anandakrishnan *et al.* 2007; Horgan *et al.* 2013). This constraining geometry provides little vertical accommodation space for the building of substantial sediment ridges through the delivery of deforming sediment to the grounding zone along a line source. Instead, sedimentary wedges of limited height, with a length of a number of kilometres in the direction of past ice flow, and sometimes extending for tens of kilometres across cross-shelf troughs, build up along the ice-sheet grounding zone (Fig. 6d) (e.g. Alley *et al.* 1989; Anderson 1999; Shipp *et al.* 1999; Mosola & Anderson 2006; Dowdeswell & Fugelli 2012; Jakobsson *et al.* 2012*a*); they are known as grounding-zone wedges (GZWs) (e.g. Powell & Domack 1995; Powell & Alley 1997; Batchelor & Dowdeswell 2015). GZWs are asymmetrical, with a steeper ice-distal face of a few degrees and a more extensive ice-proximal component of often less than one degree in surface slope – their vertical expression on the seafloor is therefore relatively subdued and they are often identified by the asymmetrical shape of the depocentre in seismic-reflection records along the axes of cross-shelf troughs. Their overall volume can be several cubic kilometres (Batchelor & Dowdeswell 2015) and, like moraine ridges, their size is dependent on the rate of supply of deforming subglacial sediments to the grounding zone and the duration of any still-stand within more general regional deglaciation. Smaller sedimentary wedges, only a few metres high and 150–500 m wide but of similar geometry to the large-volume features described above, have recently been observed in high-resolution imagery of the western Ross Sea (Simkins *et al.* 2016). By reference to the few observations available on rates of subglacial sediment delivery (e.g. Dowdeswell *et al.* 2004*a*; Alley *et al.* 2007; Christoffersen *et al.* 2010), large GZWs are calculated to form over at least decades and more probably over centuries (e.g. Ottesen *et al.* 2005*b*; Alley *et al.* 2007; Dowdeswell & Fugelli 2012; Jakobsson *et al.* 2012*a*).

### *Ice-marginal landforms: sedimentary fans*

Two main types of ice-marginal fan are present on high-latitude glacier-influenced margins; they are very different in origin and scale. The first are huge, line-sourced trough-mouth fans (TMFs) that are produced at the seaward margins of fast-flowing ice streams which reach the continental shelf edge under full-glacial conditions (e.g. Vorren & Laberg 1997; Vorren *et al.* 1998). The very large size of TMFs means that their presence can often be inferred through the identification of outward-bulging bathymetric contours on the shelf edge and continental slope within regional maps of the Arctic and Antarctic seas (Fig. 7a) (e.g. Canals *et al.* 2003; Jakobsson *et al.* 2012*b*; Arndt *et al.* 2013). These fan-shaped features are located adjacent to cross-shelf troughs that, under full-glacial conditions, held the fast-flowing ice streams which delivered large volumes of mainly unsorted diamictic debris to the shelf edge along a line source – the ice-sheet grounding zone (e.g. Dowdeswell & Siegert 1999). This debris appears to fail on the upper continental slope as glacigenic debris-flows to produce downslope mass-wasting deposits known as debrites, that are the important building blocks in TMF growth (Fig. 7a–c) (e.g. Elverhøi *et al.* 1997; Nygård *et al.* 2002). Individual debris-flow deposits are up to about 50 m thick, a few kilometres wide and extend for up to about 200 km down the slope (e.g. Laberg & Vorren 1995; Dowdeswell *et al.* 1996; King *et al.* 1996; Taylor *et al.* 2002). Delivery of deforming soft sediment to the shelf edge is clearly a full-glacial process, and the few radiocarbon dates available on the timing of glacigenic debris-flow activity support the view that the debris flows are full-glacial phenomena (e.g. Laberg & Vorren 1995; Pope *et al.* 2016). TMFs may also contain some meltwater-derived debris and turbidites, together with a hemipelagic component that normally marks slower interglacial sedimentation when ice sheets have retreated to fjord heads or onto land (e.g. Hesse *et al.* 1999; Taylor *et al.* 2002; Piper *et al.* 2007; Ó Cofaigh *et al.* 2013). Greater availability of meltwater is likely to be important in the proportion of meltwater-delivered sediment to TMFs at the termini of mid-latitude as compared with high-latitude palaeo-ice streams, where conglomerates are formed from major meltwater-discharge events (e.g. Piper *et al.* 2007, 2012, 2016).

Huge TMFs, covering areas of $10^3$–$10^5$ km$^2$ and with volumes typically of $10^4$–$10^5$ km$^3$ are by far the largest glacier-influenced landforms in the seas of the Arctic and Antarctic (Fig. 7a) (e.g. Dowdeswell *et al.* 1997, 2008; Vorren & Laberg 1997; Vorren *et al.* 1998; Canals *et al.* 2003). They are comparable in size to the largest river-fed fan systems at lower latitudes, such as those of the Amazon and Mississippi rivers (Dowdeswell *et al.* 1998, 2010), although the internal architecture and grain size of glacier-fed TMFs are very different to those of the great

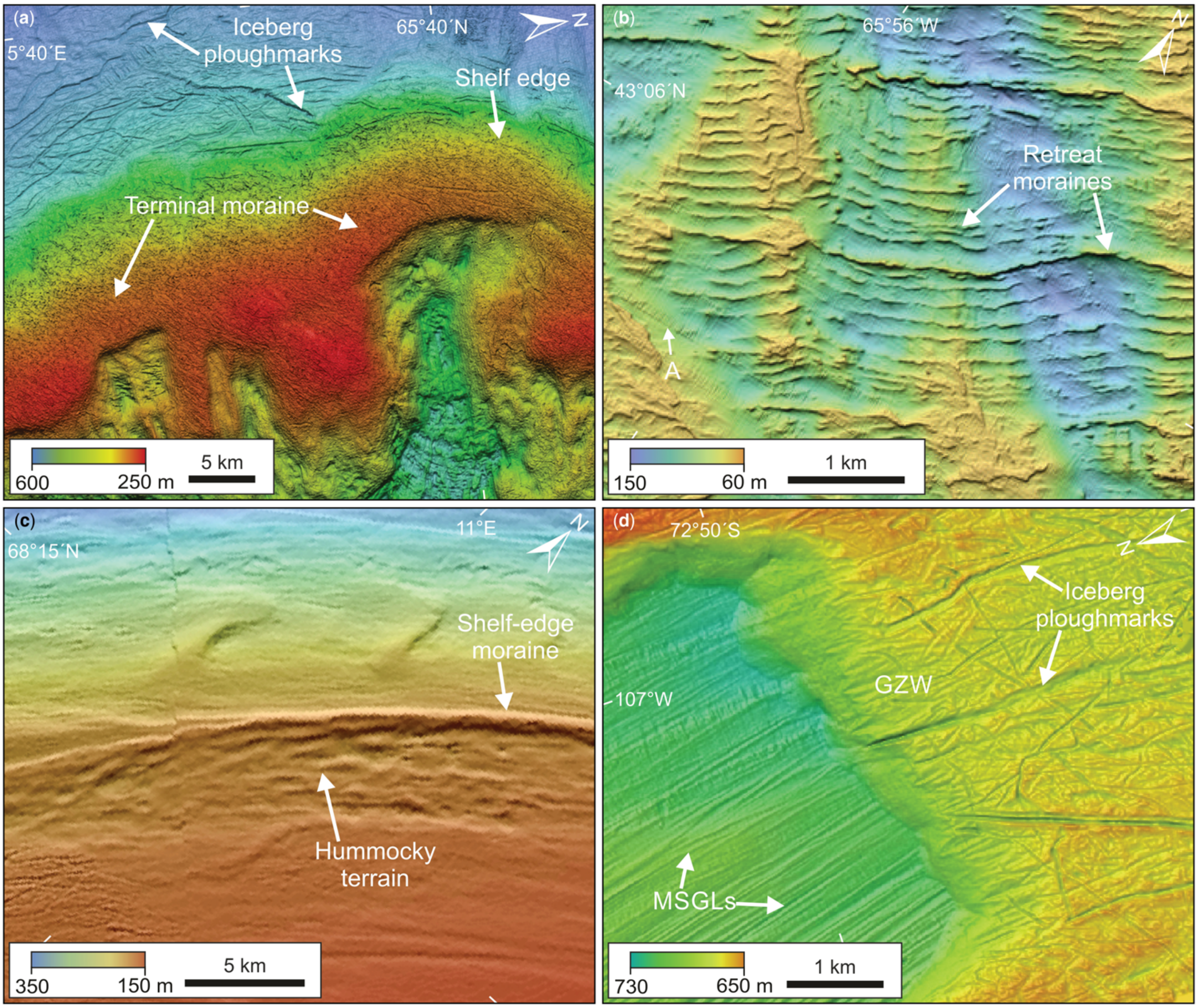


**Fig. 6.** Submarine examples of ice-marginal landforms. (**a**) The huge Skjoldryggen terminal moraines off mid-Norway (modified from Dowdeswell *et al.* 2016*g*). Some iceberg ploughmarks are visible on the ridge crest and beyond. (**b**) Transverse-to-flow De Geer retreat moraines on the Scotian Shelf, Atlantic Canada (modified from Todd 2016*b*). (**c**) Hummocky terrain and a shelf-edge moraine ridge off mid-Norway (modified from Dowdeswell *et al.* 2016*h*). (**d**) A grounding-zone wedge (GZW) on the floor of Pine Island Bay, West Antarctica (modified from Anderson & Jakobsson 2016). The GZW overrides mega-scale glacial lineations (MSGLs) and has ploughmarks on its surface.

fluvial fans, being composed largely of diamictic debris-flows rather than sorted sediments. TMFs are not present beyond all ice stream-filled troughs, however. Where the continental slope is relatively steep at greater than a few degrees, often due to long-term tectonic influences, glacigenic-sediment bypass of the upper slope appears to take place with debris being transferred via submarine channels to deep-ocean basins (Fig. 7d) (e.g. Stevenson *et al.* 2015). The slope beyond Marguerite Trough on the Antarctic Peninsula is a clear example of the presence of an ice stream without a major TMF beyond the trough mouth (e.g. Dowdeswell *et al.* 2004*b*).

A second type of fan, which is usually several orders of magnitude smaller in area and volume than TMFs, is the ice-proximal fan; such fans are produced at point sources where subglacial meltwater streams enter marine waters and deposit their sediment load as they lose energy (e.g. Pfirman & Solheim 1989; Powell 1990; Mugford & Dowdeswell 2011; Dowdeswell *et al.* 2015). These ice-proximal fans are usually up to a few cubic kilometres in volume and thin over a few kilometres in length from apex to toe (Fig. 8a). Glacimarine deposition from subglacial meltwater streams typically produces well-sorted sediments that fine with distance, from gravels and sands close to the ice-cliff portal to fine-grained sedimentation as overflowing buoyant plumes of suspended sediment progressively dominate more distally (e.g. Syvitski 1989; Powell 1990; Dowdeswell *et al.* 2015). Ice-proximal fans produced by point-sourced meltwater delivery of sediments are found at the margins of modern, Quaternary and more ancient marine- and lake-terminating glaciers (e.g. Hirst 2012; Koch & Isbell 2013), but are perhaps less common than once thought because they require an ice margin that is stable over years to decades in order to deliver the volume of debris required to build an identifiable fan (Dowdeswell *et al.* 2015). The largest ice-proximal fans are probably produced when ice-dammed subglacial or surface lakes drain rapidly to marine- or lake-terminating ice margins (e.g. Winsemann *et al.* 2009); such catastrophic events can take place over days to weeks (e.g. Russell

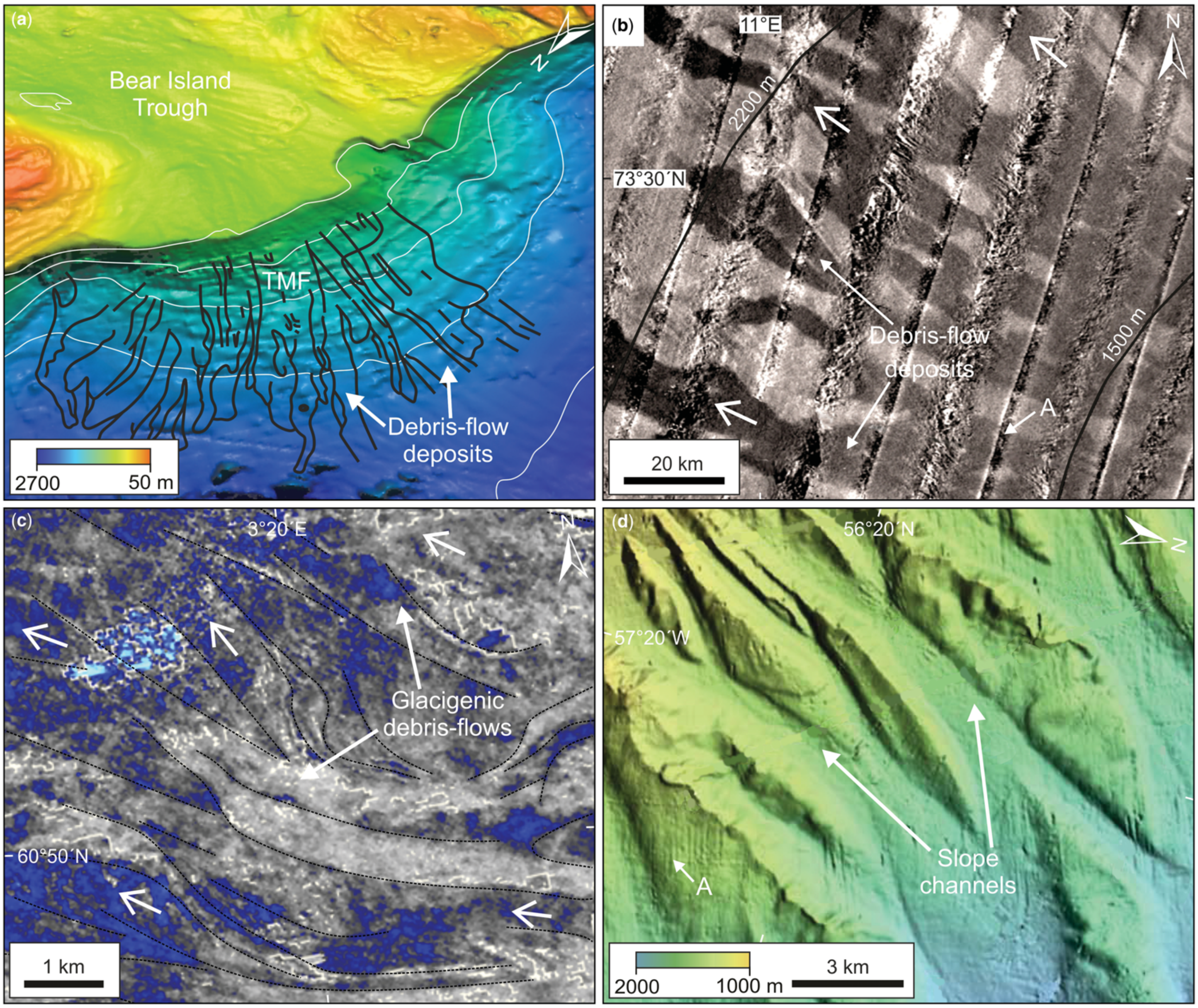


**Fig. 7.** Glacial landforms on high-latitude continental slopes. (**a**) The Bear Island TMF, Barents Sea margin, with glacigenic debris-flows on its surface (modified from Laberg & Dowdeswell 2016). (**b**) Glacigenic debris-flows on the Bear Island Fan imaged from GLORIA 6 kHz system (modified from Laberg & Dowdeswell 2016). (**c**) Buried glacigenic debris-flows in the North Sea Fan imaged on a palaeoshelf from 3D seismic-reflection data (modified from Ottesen *et al.* 2014). (**d**) Submarine channels on the Labrador slope, Atlantic Canada, offshore of a cross-shelf trough (modified from Dowdeswell *et al.* 2016*d*). A, artefacts.

*et al.* 2001, 2006), implying that decadal ice-marginal stability is not needed in these circumstances. Grading out from the toes of ice-proximal fans, and in fjord basins increasingly distal from point sources of subglacial meltwater delivery, flat-floored basins with acoustically laminated basin-fill are often found (Fig. 8b). These deposits are produced by the rain-out of fine-grained suspended sediments from turbid meltwater plumes derived from subglacial meltwater drainage to tidewater glacier margins (e.g. Ottesen & Dowdeswell 2009; Ó Cofaigh *et al.* 2016*b*).

A third form of glacier-influenced depocentre occurs at the mouths of glacier- and snowmelt-fed braided-river systems (i.e. sandurs) which are present in many deglaciated or partially glaciated Arctic valleys (e.g. Church & Gilbert 1975). Where these braided rivers reach the sea, relatively coarse-grained submarine deltas graded to sea level are typically developed beyond the shifting mouths of the braided channels (Fig. 8c) (e.g. Syvitski *et al.* 1987; Eilertsen *et al.* 2016). A number of mass-wasting processes contribute to downslope sediment transfer on such deltas (e.g. Prior *et al.* 1984).

### *Glacimarine landforms*

The most characteristic submarine landforms produced directly by ice beyond the margins of marine ice sheets are the curvilinear depressions and associated berms formed by the ploughing action of drifting iceberg and sea-ice keels where they impinge on the sedimentary seafloor (Fig. 9); these features are commonly referred to as ploughmarks, scours or furrows (e.g. Woodworth-Lynas *et al.* 1991). Iceberg keels vary in depth depending on the type of ice-sheet margin from which they are calved (Dowdeswell & Bamber 2007). The deepest-keeled icebergs tend to come from ice-stream and fast-flowing outlet glacier margins that either are grounded or have only a short floating section beyond the grounding zone. The presence of only limited floating ice shelves restricts thinning by basal ice-shelf melting, which can reach more than 10 m $a^{-1}$ (e.g. Enderlin & Howat 2013; Rignot *et al.* 2013). Ice-shelf thinning by internal creep, where the floating ice spreads out under its own weight (Sanderson 1979), is also limited where ice-shelf length is short. Large icebergs from these

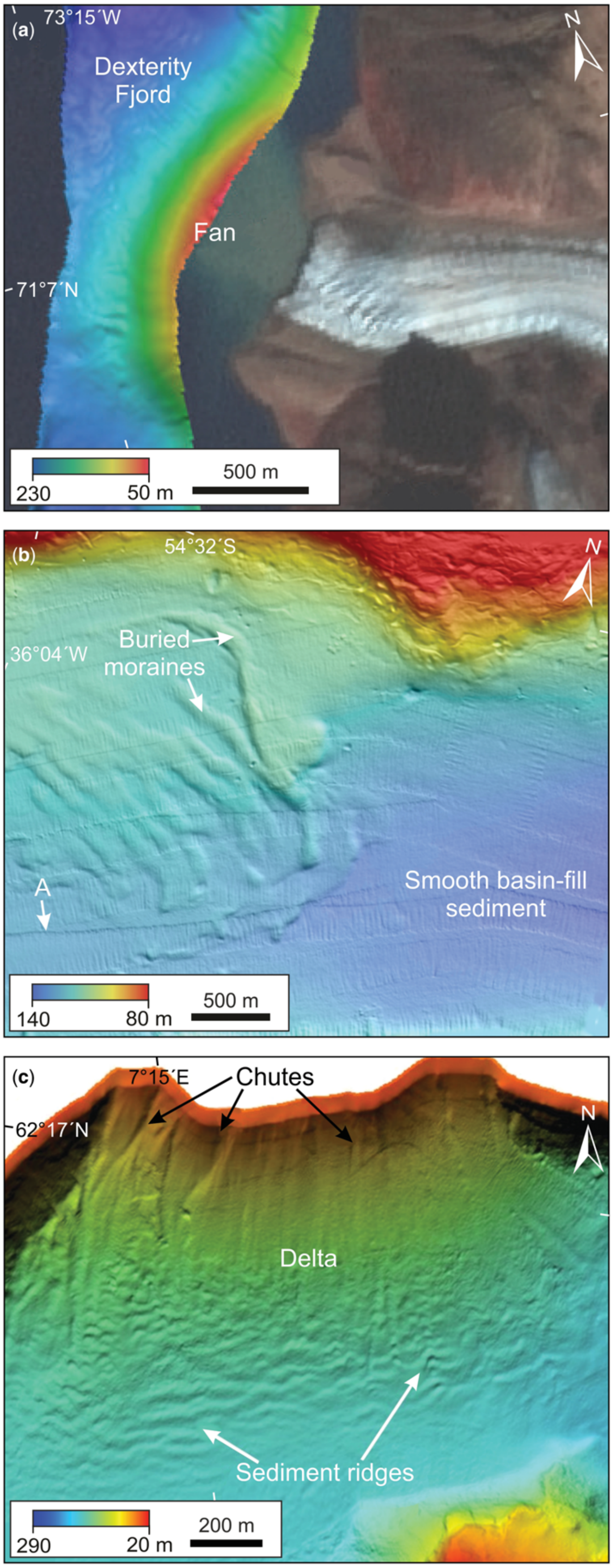


**Fig. 8.** (**a**) Ice-proximal fan and tidewater glacier in eastern Baffin Island, Canadian Arctic (modified from E. Dowdeswell *et al.* 2016*a*). (**b**) Smooth basin-fill and buried moraine ridges in Royal Bay, South Georgia (modified from Graham & Hodgson 2016). Example of an artefact is labelled A. (**c**) Fjord delta fed from a fluvial system in Valldal, Norway (modified from Eilertsen *et al.* 2016).

sources can exceed 500 m in keel depth (Barnes & Lien 1988; Dowdeswell *et al.* 1992); occasional ploughmarks have been reported from depths of 800–1000 m (e.g. Kuijpers *et al.* 2007; Jakobsson *et al.* 2010; Dowdeswell & Hogan 2016). By contrast, where large floating ice shelves of tens to hundreds of kilometres in length are present, for example, the Ross and Ronne ice shelves in Antarctica, the processes of thinning mean that calved icebergs are often no more than about 300 m in keel depth (Dowdeswell & Bamber 2007). Ice shelves of considerable thickness are also thought to have existed in the central Arctic Ocean during some past full-glacial periods and have left traces of grounding on bathymetric highs. On the Lomonosov Ridge, extending across the central Arctic Ocean, ice-shelf grounding from the penultimate glaciation (*c.* 140 ka) has left extremely parallel lineations that resemble MSGLs (Fig. 9d) (Jakobsson *et al.* 2016*c*).

It has been estimated that about half of the 500 000 km$^2$ Greenland continental shelf has been reworked by the ploughing action of iceberg keels, particularly at water depths of less than about 400 m (Brett & Zarudzki 1979; Dowdeswell *et al.* 1993, 2014; Syvitski *et al.* 2001). Chaotic patterns of buried iceberg ploughmarks have also been identified in 3D seismic cubes from, for example, the North Sea (Fig. 9c), where they have been used to identify the inception of early Quaternary ice cover reaching to sea-level (Dowdeswell & Ottesen 2013) and to reconstruct past ocean-current directions (Newton *et al.* 2016). Grounded tidewater glacier margins in fjords tend to produce large numbers of irregularly shaped small icebergs with keels of only a few tens of metres (Dowdeswell & Forsberg 1992) because the dimensions of calved icebergs are often restricted by the close density of crevasses near glacier termini. The underwater keels of sea-ice floes are also limited in depth because sea ice is usually only a few metres thick when it forms, and even multi-year floes that have undergone significant pressure in pack ice have submarine ridges of a few tens of metres at most (e.g. Wilkinson & Wadhams 2016). Large numbers of sea-ice-derived ploughmarks are, therefore, found only on shallow continental shelves such as those of the Beaufort and Laptev seas (e.g. Héquette *et al.* 1995; Ananyev *et al.* 2016) where water depths are only a few tens of metres over very large areas.

As icebergs drift through ocean waters, they melt at a rate that depends on water temperature, drift velocity and berg dimensions (e.g. Weeks & Campbell 1973; Mugford & Dowdeswell 2010). Any debris held within icebergs, especially in the quantitatively important debris-rich zone formed at the parent ice-sheet base (Dowdeswell & Murray 1990), will rain out as melting proceeds; any sediment accumulated on the surface will be dumped occasionally as icebergs fragment and overturn. The sediment deposited from melting and rain-out is known as iceberg-rafted debris (IRD) (e.g. Gilbert 1990), and has been used to identify, for example, the Heinrich Layers that form six distinct horizons of coarser-grained debris deposited within the predominantly fine-grained muds of the North Atlantic over the past 60 000 years or so (e.g. Heinrich 1988; Bond *et al.* 1992; Dowdeswell *et al.* 1995). IRD deposition influences the grain size and stratigraphy of glacimarine sediments but it does not necessarily produce distinctive submarine landforms. Nonetheless, it has been suggested that pods of relatively coarse-grained material in Quaternary lake sediments in Scotland provide evidence for so-called iceberg dump structures, which are cone-shaped features resulting from large quantities of debris being tipped off overturning icebergs (Thomas & Connell 1985). No such landforms have been identified on the high-latitude seafloor, however.

### *Other submarine landforms on high-latitude margins*

In addition to the submarine landforms produced specifically in association with ice, high-latitude continental margins are also affected by a number of processes that operate more widely in the world's oceans (Table 1). Submarine landforms produced by

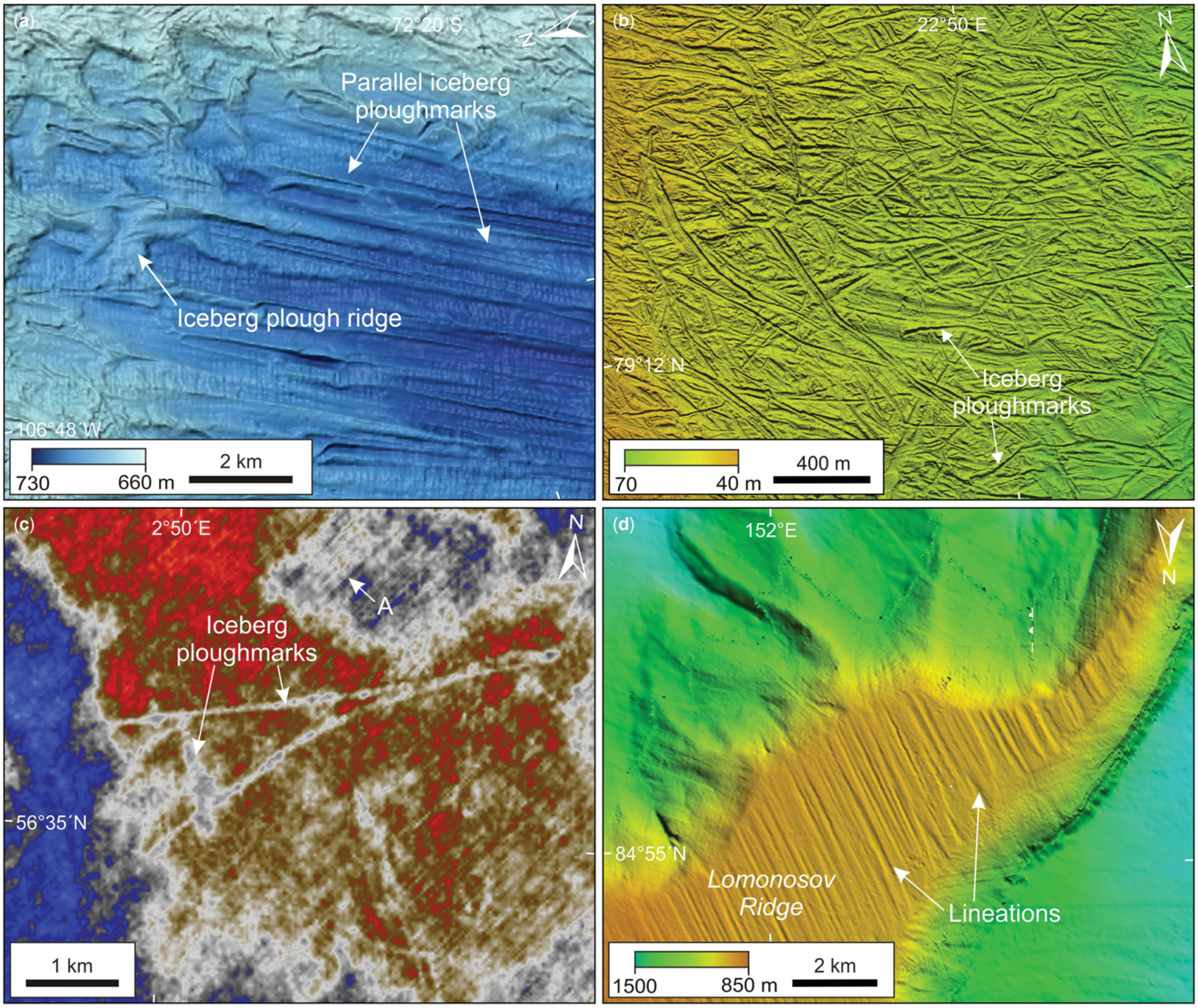


**Fig. 9.** Ice-keel ploughmarks in the glacimarine record. (**a**) Large sub-parallel iceberg ploughmarks and push-ridges in Pine Island Bay, West Antarctica (modified from Jakobsson & Anderson 2016). (**b**) Chaotic iceberg ploughmarks off Austfonna, northern Barents Sea, eastern Svalbard (courtesy of the Norwegian Hydrographic Service, permission no. 14G/754). (**c**) Iceberg ploughmarks buried about 700 m deep in the central North Sea and imaged from 3D seismic-reflection data (modified from Dowdeswell & Ottesen 2016*b*). A marks a processing artefact. (**d**) Enigmatic MSGL-like lineations on the Lomonosov Ridge, Arctic Ocean, probably formed by ice-shelf grounding (modified from Jakobsson *et al.* 2010).

a variety of mass-wasting processes include slides and debris flows, both of which occur at a number of locations and scales. Slope failures in the form of slides (Fig. 10a), where there is failure along discrete buried décollement surfaces, are present at the base of fjord walls after sediment has failed on steep slopes; these slides are often just a few hundreds of metres to a few kilometres in length (Ottesen & Dowdeswell 2009; Dowdeswell *et al.* 2016*e*). At much broader scales, but on much lower-gradient continental slopes, huge slides of thousands to tens of thousands of square kilometres in area are represented by, for example, the Storegga, Traenadjupet, Andøya and Hinlopen slides on the Norwegian and Svalbard margins (Fig. 10b) (Bugge *et al.* 1987; Dowdeswell *et al.* 2002; Vanneste *et al.* 2006; Laberg *et al.* 2007; Hogan *et al.* 2013). Although these large slides contain both glacial and hemipelagic sediments, built up on high-latitude continental slopes through glacial and interglacial periods, respectively, there appears to be limited evidence on the systematic timing of failure in relation to ice-sheet growth and decay. Current-produced contourites can act as weak layers where slope failure may take place, and can also infill former landslide depressions on the continental slope (e.g. Canals *et al.* 2004; Bryn *et al.* 2005).

Submarine debris-flows, where internal deformation as well as movement over a glide plane are important to motion, also occur at several scales and involve the downslope movement of glacigenic debris. Examples range from debris-flow deposits on the distal slopes of moraine ridges (Fig. 10c) (e.g. Ottesen & Dowdeswell 2006), of hundreds to thousands of metres in run-out distance, to debrites that are several hundred kilometres long on the continental slope and are described above as the building blocks of TMFs on the slope beyond cross-shelf troughs (Fig. 7a–c) (e.g. Laberg & Vorren 1995; King *et al.* 1996; Taylor *et al.* 2002).

Submarine gullies, channels and canyons are also common features on continental slopes and outer shelves globally, although large erosional canyons appear to occur less frequently on glacier-influenced margins probably because of the relatively rapid rate of sediment delivery during successive full-glacial periods and

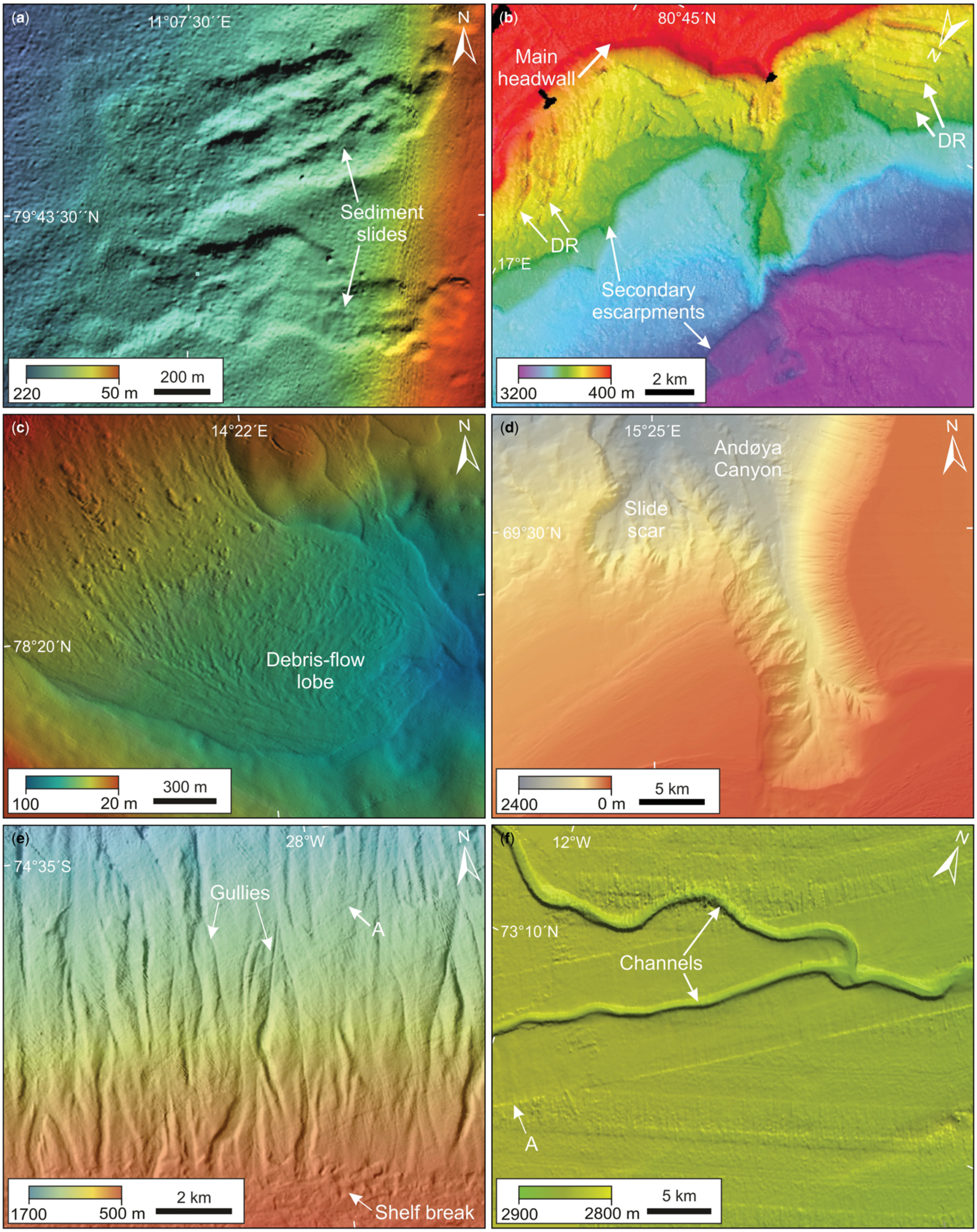


**Fig. 10.** Other features on the seafloor of high-latitude fjords, shelves and slopes. (**a**) Sediment slides off the fjord walls of Smeerenburgfjorden, NW Svalbard (modified from Dowdeswell *et al.* 2016*e*). (**b**) The upper part of the huge Hinlopen Slide with the outer shelf north of Svalbard in red (modified from Hogan *et al.* 2013). DR are detached sediment ridges. (**c**) Debris-flow lobes on the distal slope of a terminal-moraine ridge, Borebukta, Spitsbergen (modified from Dowdeswell *et al.* 2016*f* ). (**d**) Andøya Canyon, Barents Sea margin (modified from Laberg *et al.* 2016). (**e**) Gullies on the slope offshore of the Weddell Sea shelf edge, Antarctica (modified from Gales *et al.* 2016). (**f**) The confluence of two turbidite channels on the abyssal seafloor of the Greenland Basin (modified from García *et al.* 2016). Features labelled A in panels (e) and (f) are artefacts.

the consequent progradation of the continental shelf. A number of examples of canyons do exist on high-latitude margins, however (Fig. 10d) (e.g. Scholl *et al.* 1970; Fader *et al.* 1998; Taylor *et al.* 2000; Amblas *et al.* 2006; Laberg *et al.* 2007; Rise *et al.* 2013; Cameron *et al.* 2016; E. Dowdeswell *et al.* 2016*b*), sometimes beyond continental shelves that have received relatively restricted volumes of glacial sediment due to their inter-ice stream locations between fast-flowing ice. The development of these canyons, presumably over the past few million years given their large dimensions, is thought to be related to periods when lower sea-level enabled fluvial activity to extend to the outer shelf and when meltwater erosion took place linked to full-glacial turbidity-current activity during the Quaternary, as well as to headward erosion by retrogressive mass failures (Harris & Whiteway 2011).

Relatively narrow and often V-shaped gullies have been reported on many upper slope areas of high-latitude margins, with some incised back into the outermost shelf (Fig. 10e) (e.g. Noormets *et al.* 2009; Gales *et al.* 2012, 2013). Their formation has often been ascribed to dense meltwater flows from full-glacial ice sheets extending to the shelf edge, which can operate in combination with retrogressive slope failure (e.g. Lowe & Anderson 2002; Dowdeswell *et al.* 2006; Gales *et al.* 2012). It is thought that deeper gullies, which can be tens of metres in depth, may be linked to continuing development over successive glacial cycles.

On the continental slope, submarine channels are sometimes found immediately below gullies on the uppermost slope, whereas in other cases the two types of conduits for downslope mass transfer appear disjunct, as observed on parts of the Labrador margin (e.g. Dowdeswell *et al.* 2016*d*). Turbidity currents of dense water may form from the relatively rapid delivery of meltwater and sediment to the upper slope in association with full-glacial ice at the shelf break, and also because of cooling of surface waters and subsequent sinking which can involve brine rejection during sea-ice formation (e.g. Noormets *et al.* 2009; Gales *et al.* 2012); they have a demonstrated capability to entrain significant sediment loads. In addition, when ice is present, meltwater delivery from ice-sheet margins and dense water from sea-ice formation are likely to be quasi-continuous processes at the shelf edge, with possible implications for time-transgressive submarine-channel formation (Hodgson *et al.* 2016). Both downslope-moving dense-water and debris flows occurring on the upper slope are inferred to transform into turbidity currents that may form relatively wide channels on the mid- and lower-continental slope, which may continue for hundreds of kilometres across the continental rise and onto very low-gradient abyssal plains in the deep ocean. At their distal ends, sandy channel-mouth lobes have been recorded in such deep-ocean basins (e.g. Dowdeswell *et al.* 2002; Ó Cofaigh *et al.* 2004; García *et al.* 2012, 2016). Turbidity-current channels can be up to hundreds of kilometres long and several kilometres wide, reaching over 100 m deep, with extensive levees beyond formed by flow-stripping and overspill (Fig. 10f) (e.g. Chough & Hesse 1976; Hesse *et al.* 1987; Mienert *et al.* 1993; Wilken & Mienert 2006; García *et al.* 2016). Dating of calcareous microfossils in the levees of the major submarine-channel system of the Greenland Basin, offshore of East Greenland, demonstrates that most activity is full-glacial, suggesting that the presence of ice at or close to the shelf edge is important to the supply of debris in relatively dense downslope flows (Ó Cofaigh *et al.* 2004; García *et al.* 2012, 2016). Where submarine channels are located beyond ice streams on relatively steep continental slopes, they typically become efficient bypass systems for the delivery of debris to deep-ocean basins (e.g. Stevenson *et al.* 2015), restricting the building of well-developed TMFs; the channel systems off the northern Antarctic Peninsula Pacific margin are good examples of this setting (Rebesco *et al.* 1996; Dowdeswell *et al.* 2004*b*; Amblas *et al.* 2006).

Sinuous turbidity-current channels are also common on fjord floors, incised into laminated basin-fill that has built up during and subsequent to deglaciation (Fig. 11a). In the relatively mild settings of, for example, SE Alaskan, British Columbian and Patagonian fjords, where much sediment is delivered through annual ice- and snow-melt and sedimentation rates are high at centimetres or more per year (Powell & Molnia 1989; Seramur *et al.* 1997; Cowan *et al.* 2010), turbidity currents are often activated by strong spring runoff, high sediment loads and delta-front failure (e.g. Hughes Clark 2016). Such regular activity implies that turbidity-current channels are often present in the otherwise relatively undifferentiated Holocene sediments of many fjord floors where meltwater and sediment are delivered seasonally from glacier-fed braided-river systems (e.g. Syvitski *et al.* 1987; Carlson *et al.* 1989; Dowdeswell & Vásquez 2013; Stacey & Hill 2016).

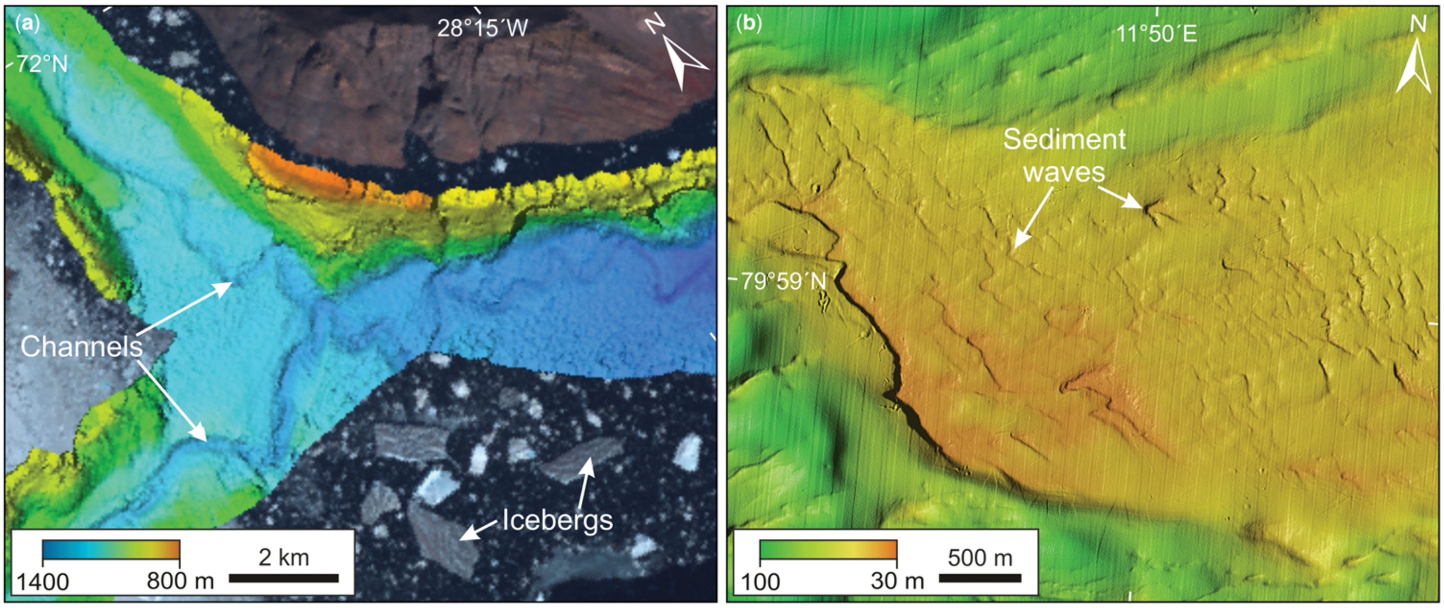


**Fig. 11.** (**a**) Turbidite channels on the floor of Nordvestfjorden, inner Scoresby Sund, East Greenland (modified from Dowdeswell *et al.* 2016*a*). (**b**) Sediment waves on a shallow moraine ridge, Raudfjorden, NW Svalbard (modified from Dowdeswell & Ottesen 2016*c*). Parallel stripes, which are processing artefacts, trend roughly north–south in the image.

Ocean current-related processes also affect high-latitude margins. Between the well-developed submarine-channel systems west of the Antarctic Peninsula, a series of up to 1 km-thick drifts of sorted sediments have built up since the late Neogene when the finer fraction of downslope turbidity currents was first mobilized in, and then deposited from, SW-flowing along-slope contour currents at about 2000–3500 m depth (e.g. Rebesco *et al.* 1996; Pudsey & Camerlenghi 1998; Amblas *et al.* 2006). At finer spatial and temporal scales, shelf and fjord sediments may be modified and sometimes buried by bedforms such as sand waves and current ripples associated with relatively strong seafloor currents above a few tens of centimetres per second (Fig. 11b). Where relatively strong currents of greater than about 0.5 m $s^{-1}$ affect unsorted diamicts on polar shelves, winnowing of fine material can lead to the formation of gravel pavements which armour the seafloor against further erosion.

In addition, the seafloor, including that at high latitudes, may be reworked by several further processes. These include the venting of gas or water from sub-seafloor sources due to, for example, loading and unloading of shelf sediments by ice and water. Such venting produces crater-like pockmarks on the seafloor that are typically several tens of metres wide and a few metres deep (Fig. 12a) and can also be identified in the stratigraphic record by chimney-like disturbances of otherwise acoustically laminated sediments (e.g. Chand *et al.* 2016; Dowdeswell *et al.* 2016*c*). Finally, human activity may also mark and scar the seafloor, for example, through the action of trawling by fishing vessels on Arctic shelves and slopes (Fig. 12b) (e.g. Thorsnes *et al.* 2016), the emplacement of submarine pipelines (Fig. 12c), cables and well-heads, and the presence of dumped waste from industrial activity at the edge of fjords (e.g. Shaw & Potter 2016).

## Volume and time frame for formation of submarine glacial landforms

The volume of a number of the submarine glacial and related landforms discussed above, together with the estimated time that they take to build up, is plotted in Figure 13. Marine-geophysical observations of area and thickness are available to characterize the approximate volume of each submarine landform type, but less control is available on the time taken for such landforms to develop, so this axis is therefore less closely constrained in Figure 13.

The volume of submarine glacial landforms ranges over ten orders of magnitude, with individual berms at iceberg-ploughmark margins at $10^{-4}$ $km^3$, and MSGLs and small transverse or De Geer moraines among the lowest in volume at $10^{-2}$–$10^{-3}$ $km^3$, contrasting with huge TMFs with sedimentary depocentres of $10^4$–$10^5$ $km^3$. Most GZWs are of $10^1$–$10^2$ $km^3$ in volume (Batchelor & Dowdeswell 2015), but some recently observed smaller sediment wedges of similar geometry are only about $10^{-2}$ $km^3$ (Shipp *et al.* 2002; Simkins *et al.* 2016). In terms of timescales for the formation of specific submarine landforms, the range is some seven orders of magnitude, from a single year or less to millions of years. Some small transverse moraines may form in just one season when minor winter readvances of tidewater glaciers push up small ridges before more major retreat each summer. Ice-proximal fans typically form over a few decades and larger GZWs probably over decades to a few centuries. Major TMFs that cause seaward progradation of high-latitude shelves, produced by rapid ice-stream delivery of diamictic sediment during successive glacial cycles, form on timescales of hundreds of thousands of years to the whole Quaternary and, in some Antarctic cases, through much of the Late Cenozoic. By contrast, some landforms are produced almost instantaneously or over hours to days; iceberg grounding pits and ploughmarks are clear examples.

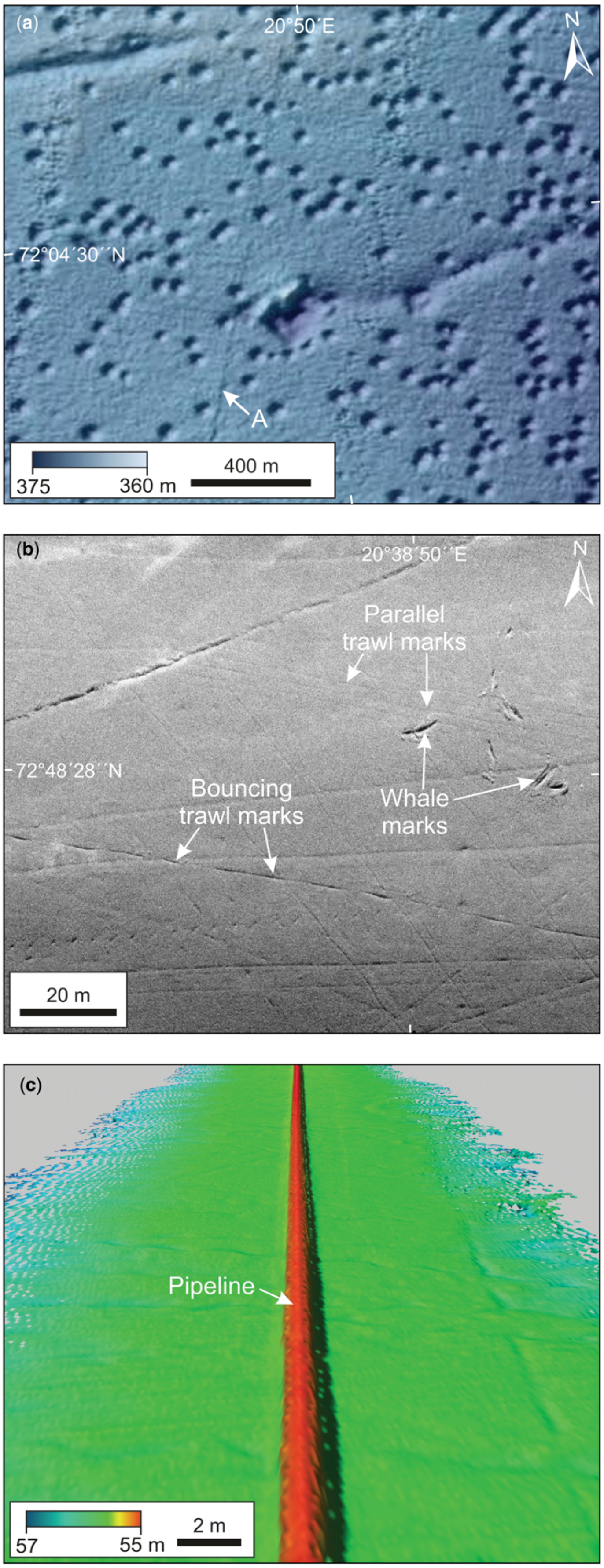


**Fig. 12.** Disturbance of seafloor sediments on high-latitude shelves. (**a**) Pockmarks in the SW Barents Sea (modified from Chand *et al.* 2016). (**b**) Trawl marks and possible whale-feeding marks in the Barents Sea (modified from Thorsnes *et al.* 2016). (**c**) Pipeline (1.5 m wide) crossing an otherwise flat area of the Baltic Sea floor (courtesy of UK Seafloor Mapping Ltd.).

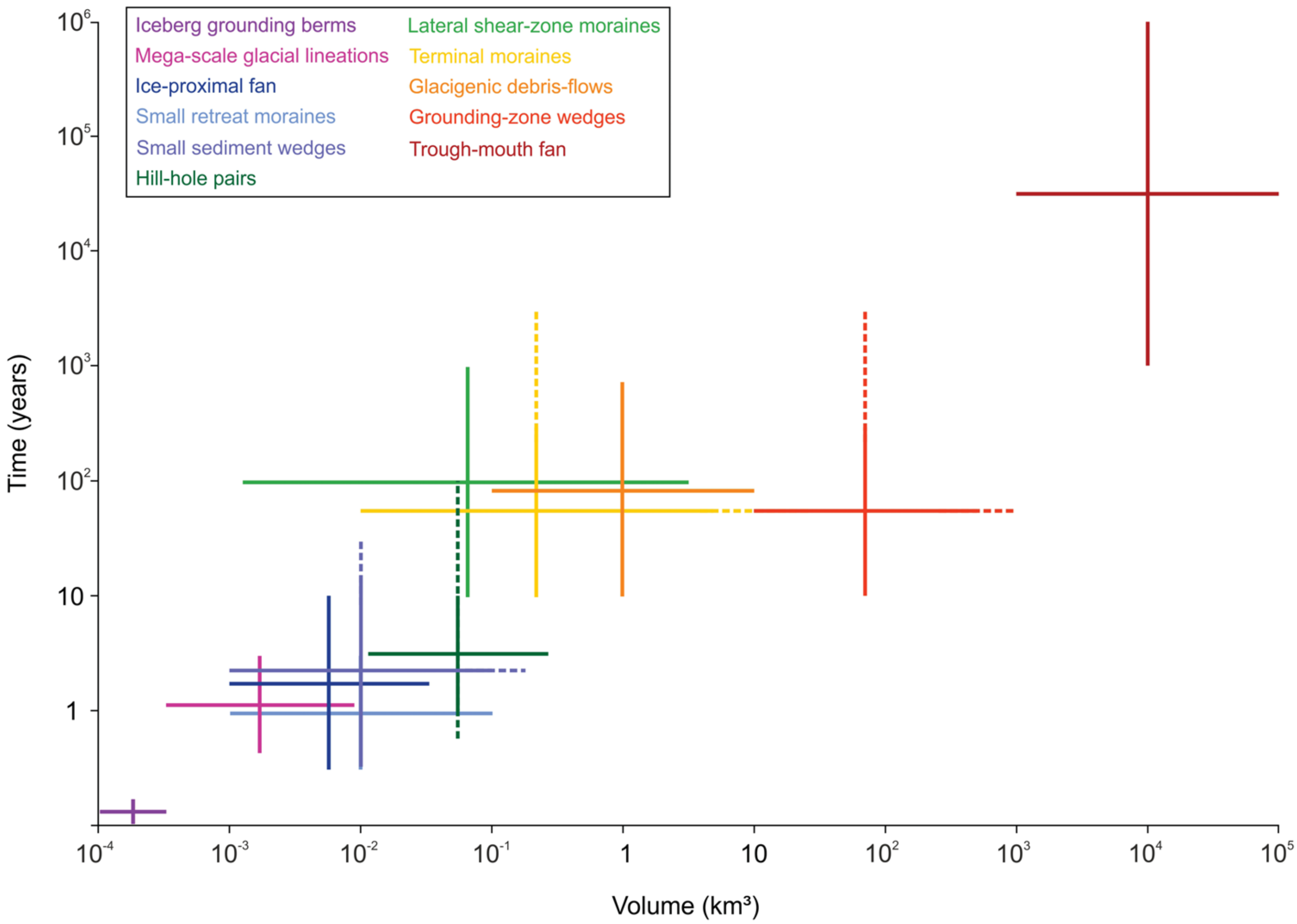


**Fig. 13.** Graph showing the estimated range in volume of a number of submarine glacial landforms and the approximate timescales over which they build up. Note the order-of-magnitude scales.

The volume of specific glacial and related submarine landforms, and the time needed for their formation, scale together; this is illustrated in Figure 13. The relationship is explained, in part at least, because the development of many different subglacial and ice-proximal landforms is related to the rate of transport and delivery of deformable till at the base of glaciers and ice sheets. Where 'catastrophic processes', such as iceberg-calving events or outburst floods of meltwater, are involved, the time frame is obviously very short, but the volumes of sediment deposited or reworked are usually less than about a cubic kilometre and often only small fractions of that volume. More broadly, it is a general characteristic of both stratigraphic units and landforms on continental margins that space and time scale together (Nittrouer & Kravitz 1996). The same applies to many geomorphic systems, for example, those involving fluvial and aeolian processes and landforms (Schumm & Lichty 1965; Dodds & Rothman 2000).

## Discussion: submarine glacial landforms and ice-sheet reconstruction

### *Past ice-sheet extent and flow direction*

At its simplest, the occurrence in a given location of submarine landforms that can be linked unequivocally to glacial activity indicates the former presence of ice. Seafloor landforms that are known to have formed subglacially or ice-marginally therefore provide clear evidence of the past extent of grounded ice sheets. This is particularly important on some high-latitude shelves where, until recently, there have been few geophysical and geological datasets available. The continental shelf of Northeast Greenland is a good example, in which swath-bathymetric evidence of a combination of subglacially produced MSGLs and transverse-to-flow ice-marginal retreat moraines demonstrated for the first time that ice of probable LGM age had covered much of the 250 km-wide shelf (e.g. Evans *et al.* 2009; Winkelmann *et al.* 2010; Arndt *et al.* 2015). This compelling submarine geomorphic evidence helped replace an earlier view, based on limited terrestrial Quaternary evidence from the Northeast Greenland coast, that LGM ice had not been much more extensive there than during interglacials (e.g. Funder & Hansen 1996). On the Antarctic, Greenland, Norwegian, eastern Canadian and British shelves, for example, the presence of streamlined subglacial landforms in particular has also shown that full-glacial ice extended to the shelf edge in many areas, especially where cross-shelf troughs were present (e.g. Anderson *et al.* 2002; Ottesen *et al.* 2005*a*, *b*; Todd *et al.* 2007; Bradwell *et al.* 2008; Livingstone *et al.* 2012; Dowdeswell *et al.* 2014). Similarly, because ice-marginal retreat moraines and GZWs are usually orientated transverse to flow, the former position of the ice-sheet terminus can also be mapped (e.g. Mosola & Anderson 2006; Hogan *et al.* 2010).

In addition to simple presence or absence, the fact that many subglacial landforms are elongate, with long axes orientated in the direction of past ice flow, means that flowlines within past ice sheets can be reconstructed (Fig. 14). Taking this further, where subglacial landforms in particular have low elongation

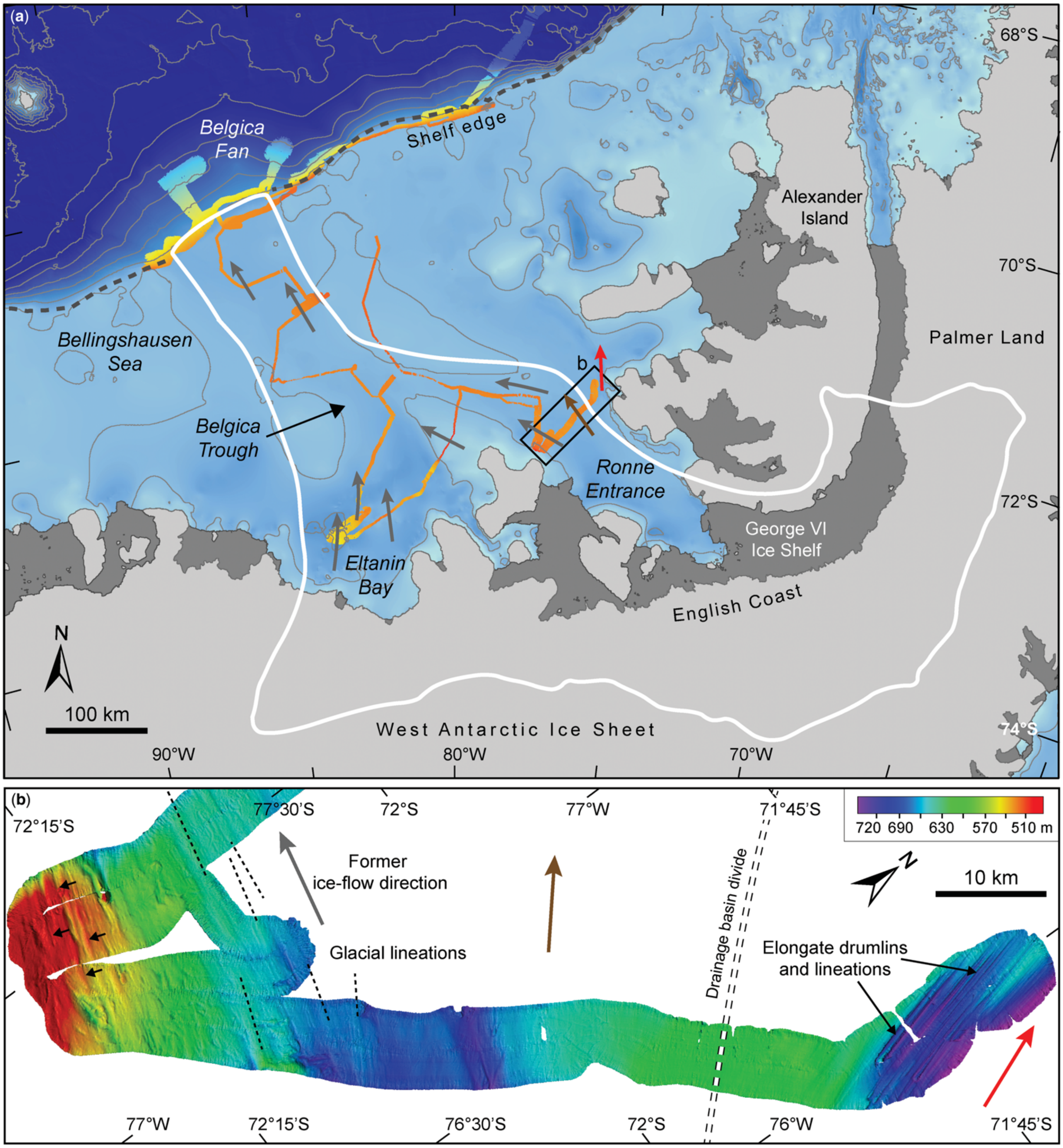


**Fig. 14.** (**a**) Ice-flow directions (indicated by arrows and based on the orientation of streamlined subglacial landforms in the coloured multibeam data) and the estimated size of the Quaternary basin (outlined in white and about 220 000 km$^2$ in area) draining full-glacial ice into the Belgica Trough area of the Bellingshausen Sea (modified from Ó Cofaigh *et al.* 2005*b*). The modern ice sheet is shaded light grey and floating ice shelves are dark grey. (**b**) Multibeam data from the inner Bellingshausen Sea and Ronne Entrance showing the orientation of streamlined subglacial landforms (modified from Ó Cofaigh *et al.* 2005*b*). Small black arrows point to crudely streamlined bedforms and thin dashed lines are streamlined glacial lineations. Larger dark grey, red and brown arrows indicate former ice-flow directions and correspond to arrows of the same colour in (a).

ratios, often associated with a blunter up-stream face and/or a more tapering sedimentary tail, for example drumlins, crag-and-tails and even the outer 'wings' of some hill–hole pairs, the sense of flow direction can be established. The mapping of past ice flowlines, in turn, enables the approximate outlines of former ice-sheet drainage basins to be defined (Fig. 14) (e.g. Ottesen *et al.* 2005*a*; Ó Cofaigh *et al.* 2005*b*).

Below the seafloor, too, the mapping of similar streamlined subglacial landforms on buried high-latitude shelves imaged in 3D seismic-reflection cubes has confirmed the former presence of grounded ice sheets during a number of Quaternary glaciations (Dowdeswell *et al.* 2006, 2007; Andreassen *et al.* 2007). Highly elongate MSGLs have even been identified on high-resolution satellite imagery of North Africa, where overlying sediments have been eroded and long-buried Late Ordovician glacier-influenced shelf areas now crop out at the surface (Moreau *et al.* 2005).

Whereas subglacial and ice-marginal landforms indicate the former presence of ice and its grounded terminus position, buried glacimarine landforms such as iceberg ploughmarks and glacigenic debrites, the latter produced by the failure of glacier-derived sediments beyond the former ice margin on the upper continental slope, indicate proximity to past ice sheets. Ploughmarks on buried continental shelves in 3D seismic-reflection data (Fig. 9c) imply that nearby ice-sheet margins must have terminated in marine waters in order to calve icebergs, and glacigenic debris-flow deposits suggest the former presence of ice in trough mouths at the shelf break (e.g. Dowdeswell *et al.* 2007; Dowdeswell & Ottesen 2013; Ottesen *et al.* 2014). The orientation of ploughmarks also yields information about past iceberg drift tracks and ocean-current directions (e.g. Newton *et al.* 2016).

### *Ice-flow dynamics: fast- and slow-flowing ice*

The distribution and nature of submarine glacial landforms can also be used to infer aspects of the past dynamics of ice sheets. Streamlined subglacial landforms have been observed forming beneath modern fast-flowing ice streams (King *et al.* 2007, 2009; Smith & Murray 2009), and MSGLs in particular have long been regarded as indicators of the presence of Quaternary ice streams in many high-latitude cross-shelf troughs (e.g. Clark 1993; Ottesen *et al.* 2005*a*, *b*). Indeed, the amount of elongation of some streamlined subglacial landforms has been linked directly to past ice-flow velocities (Stokes & Clark 2002*b*). The locations of ice streams are important to the understanding of past ice sheets, in that ice streams are the main conduits of mass loss from ice-sheet interiors, responsible for draining over 80% of the mass from the modern Antarctic and Greenland ice sheets (e.g. Rignot & Kanagaratnam 2006; Rignot *et al.* 2011). The presence of cross-shelf troughs also implies erosion by past ice streams, especially where underlying seismic reflectors have been truncated, and provides further evidence used in the reconstruction of past ice-stream locations (e.g. Batchelor & Dowdeswell 2014). The locations of former ice streams and cross-shelf troughs therefore provides a simple but potentially robust test for numerical modelling of past ice sheets in terms of whether such models can reproduce the mapped distribution of palaeo-ice streams.

Cross-shelf troughs, tens of kilometres in width and hundreds of metres in depth, are relatively large landforms (Fig. 15a). Their locations are therefore likely to indicate the positions of fast-flowing ice over at least one, and probably several, glacial cycles – major TMFs are also found beyond the mouth of many of these troughs (Fig. 15b) (Batchelor & Dowdeswell 2014). By contrast, MSGLs, with a surface expression of just a few metres, usually form over much shorter periods. Thus, the cross-cutting of sometimes several generations of MSGLs and other landforms provides evidence of a further element of past ice dynamics, specifically the switching of flow directions during ice retreat across continental shelves during the most recent deglaciation (e.g. Andreassen *et al.* 2008; Greenwood *et al.* 2012). The changing pattern of MSGLs in the inner Ross Sea around McMurdo Sound presents a clear example of such flow complexities, along with transverse-to-flow moraine ridges (Fig. 16). Using 3D seismic-reflection cubes, and at longer timescales, flow switching can also be seen between successive full-glacial periods using the changing orientations of progressively more deeply buried sets of MSGLs and the shifting position of cross-shelf troughs (e.g. Dowdeswell *et al.* 2006; Sarkar *et al.* 2011), with implications for the longer-term evolution of high-latitude continental margins over the whole Quaternary.

The assemblages of submarine landforms produced beneath ice streams also contrast greatly with the landform suites on the shallower inter-ice stream banks that typically separate one ice stream from the next along high-latitude margins (Ottesen & Dowdeswell 2009; Klages *et al.* 2013). On both the Norwegian and Antarctic shelves, cross-shelf troughs contain streamlined sedimentary landforms that typically parallel the direction of past ice flow (Fig. 17a), whereas shallower banks display landforms that are orientated mainly transverse to flow, generally in the form of moraine ridges of varying length and volume that mark the relatively slow retreat of grounded ice (Fig. 17b). Hill–hole pairs, interpreted to result from debris incorporation by basal freezing (the hole) and subsequent melting and release (the hill) in the presence of relatively thin ice, are also found on shallow banks between past ice streams (Ottesen *et al.* 2005*a*, *b*; Klages *et al.* 2013, 2016; Dowdeswell *et al.* 2016h). In addition, lateral shear-margin moraines may mark the edges of past ice streams and their interface with slower-moving ice to either side (Fig. 2f) (Stokes & Clark 2002*a*; Batchelor & Dowdeswell 2016).

An apparent exception to this morphological differentiation between ice-stream troughs and intervening banks on the basis of parallel- or transverse-to-flow landforms is the presence of major transverse-to-flow depocentres (GZWs) in many cross-shelf troughs (e.g. Mosola & Anderson 2006; Batchelor & Dowdeswell 2015). However, the GZWs, which mark stillstands of decades to centuries in ice-stream retreat, are often streamlined by MSGLs on their upper surfaces and also overlie MSGLs, implying that fast ice-stream flow continued during their formation. Indeed, their often large volume and likely limited time of formation require fast ice flow in order to supply basal-deforming sediment at a sufficient rate to allow them to develop.

### *Ice-sheet retreat rates*

The nature and distribution of submarine landforms on high-latitude continental shelves can also yield information on the style and rate of ice-terminus retreat during regional deglaciation (e.g. Mosola & Anderson 2006; Dowdeswell *et al.* 2008; Ó Cofaigh *et al.* 2008; Greenwood *et al.* 2012). Three distinctive types of deglacial ice retreat are inferred. First, MSGLs are typically present in many high-latitude cross-shelf troughs, indicating the former presence of fast-flowing ice streams. Where the MSGLs are preserved undisturbed along all or part of a trough, this provides an indication that deglacial ice retreat was rapid, presumably related to a combination of thinning and/or sea-level rise that stimulated ungrounding and collapse of all or part of the ice stream (e.g. Canals *et al.* 2000, 2016; Willmott *et al.* 2003). Secondly, where MSGLs are overprinted by one or a series of GZWs, this suggests rapid episodic or punctuated retreat between halts of decades to centuries at pinning points that probably relate, in part at least, to bathymetric highs and/or constrictions in trough width (e.g. Jamieson *et al.* 2012; Canals *et al.* 2016). Thirdly, where series of small transverse-to-flow retreat ridges are present in cross-shelf troughs, this indicates the slow retreat of a grounded ice margin; suites of sub-parallel retreat moraines can number hundreds. Given that such moraines may form regularly, and sometimes annually, these suites of ridges may indicate slow retreat over hundreds of years. Examples include Bellsund, west of Spitsbergen, and many Arctic fjords where tens to hundreds of such moraines are present (e.g. Dowdeswell *et al.* 2008).

As a variant of this slower retreat, small sedimentary wedges observed on side-scan sonar images by Shipp *et al.* (1999, 2002), and imaged at high resolution by Simkins *et al.* (2016), may represent year on year, or at least regular, retreat where vertical accommodation space is limited, possibly beneath short sections

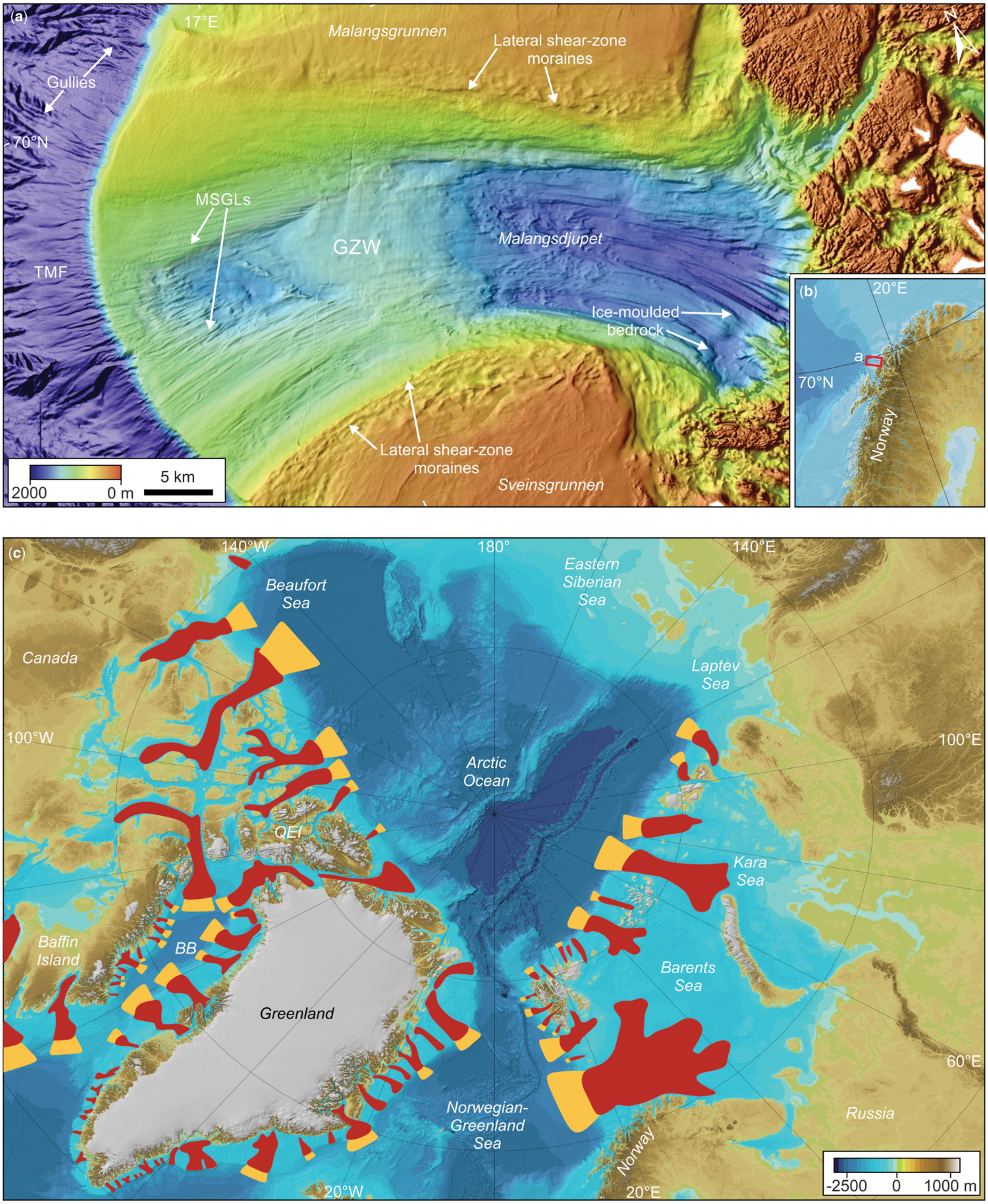


**Fig. 15.** (**a**) Multibeam-bathymetric image of Malangsdjupet cross-shelf trough on the Norwegian shelf, showing submarine landforms in the trough and on the adjacent shallower banks (image provided by the MAREANO project). (**b**) Location map (red box; map from IBCAO v. 3.0). (**c**) Distribution of major cross-shelf troughs (red) and TMFs (orange) in the High Arctic (modified from Batchelor & Dowdeswell 2014). BB is Baffin Bay, QEI are Queen Elizabeth Islands.

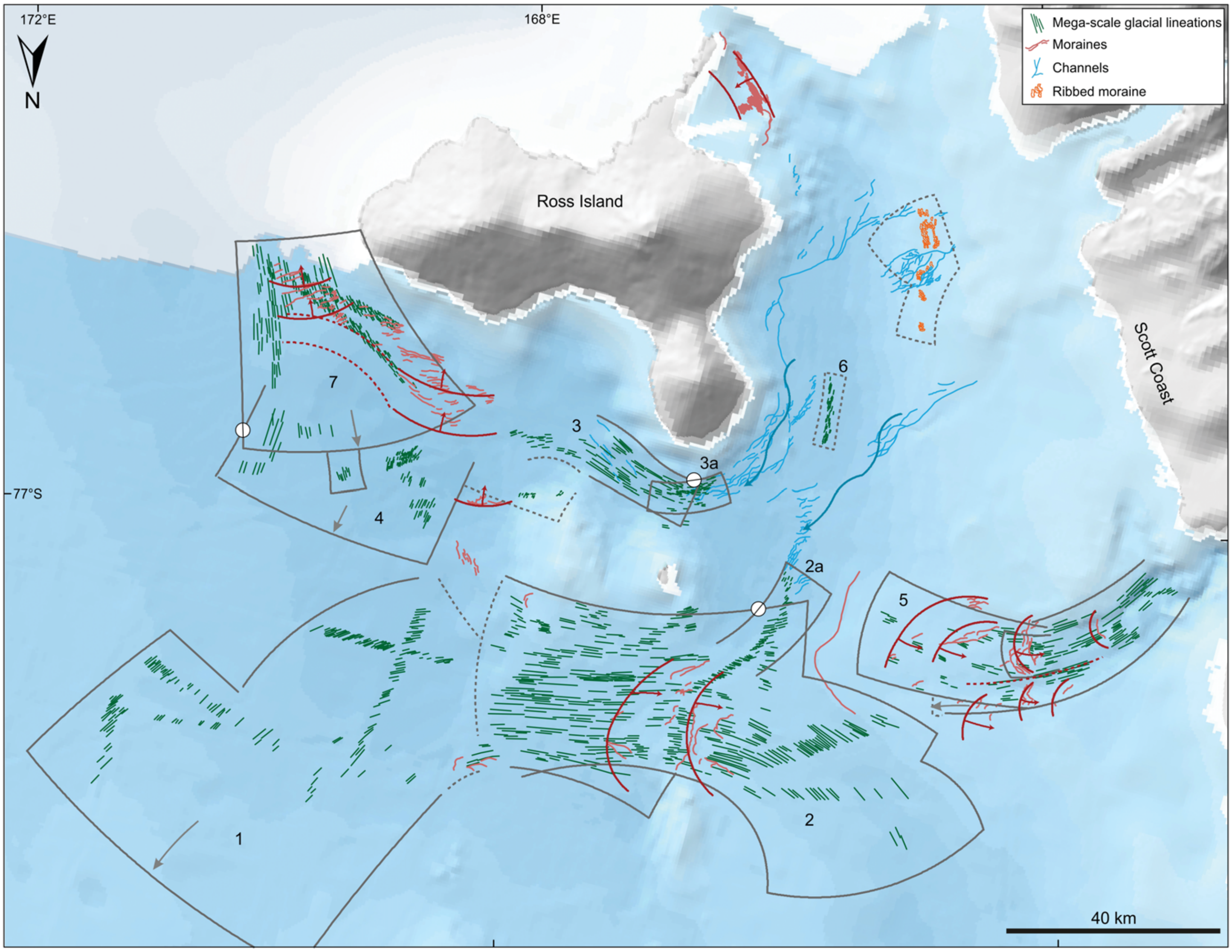


**Fig. 16.** The complexity of past ice flow shown by cross-cutting flow sets of MSGLs and other glacial landforms mapped from multibeam data in the McMurdo Sound area of the Ross Sea, Antarctica (modified from Greenwood *et al.* 2012). Sets of flow features (MSGLs parallel to ice flow and moraines transverse to flow) are indicated by grey boxes which are labelled from 1 (oldest in relative age) to 7 (youngest). Grey arrows indicate the ice-flow direction of each flow-set of MSGLs. Red arrows indicate ice-retreat direction from moraine-ridge orientations. Cross-cutting relationships are shown in white circles.

of floating ice. Ice-sheet retreat rates derived assuming the ridges are deposited annually are 40–100 m a$^{-1}$ in the western Ross Sea. This estimate is consistent with retreat of 370 km along the trough, radiocarbon-dated to the interval 14–6.4 ka BP, giving a mean retreat rate that is presumably relatively uniform at *c.* 50 m a$^{-1}$ (Shipp *et al.* 2002).

In a similar way to the locations of past ice streams on palaeo-shelves, geological evidence of the rapid, episodic or slow deglacial retreat of ice-sheet margins across continental shelves provides a useful independent test of the ability of ice-sheet numerical models to reconstruct the dynamics of past ice flow (e.g. Dowdeswell *et al.* 2008).

### *Conditions at the ice-bed interface*

The beds of modern ice sheets, and the ice streams that drain the bulk of their mass to the ocean, are overlain by several kilometres of ice, making high-resolution geophysical investigations of the ice-bed interface difficult. By contrast, well-preserved submarine landforms and sediments on formerly ice-covered continental shelves provide a window onto this interface that is much easier to study using swath-bathymetric and seismic-reflection equipment deployed from research vessels. The sedimentological and mechanical properties of deforming sediments at modern ice-stream beds have been sampled directly in only a very small number of access holes through the ice (e.g. Engelhardt *et al.* 1990; Engelhardt & Kamb 1997) and the amount of seismic-reflection data available from modern ice sheets is severely limited (e.g. Blankenship *et al.* 1987).

The observation using seismic-reflection methods that up to 6 m of deformable till was present beneath Ice Stream B (now named Whillans Ice Stream), was seminal in establishing that soft sediment rather than hard-rock beds beneath ice streams were a vital component of their fast motion of kilometres per year (Alley *et al.* 1986; Blankenship *et al.* 1987); this till deforms rapidly, especially when water pressures are high and the sediment becomes dilatant, with failures often taking place along individual shear zones within the till at any one time rather than deformation occurring simultaneously over the entire thickness of the unit (e.g. Iverson *et al.* 1996). Direct observations on the spatial extent of such deforming beds, their composition and rheology beneath modern ice streams remain very restricted, although several important studies using ice-penetrating radar and seismic-reflection methods have provided new insights (e.g. King *et al.* 2007, 2009; Smith & Murray 2009; Horgan *et al.* 2013). By contrast, the extent and thickness of deformation till has been mapped

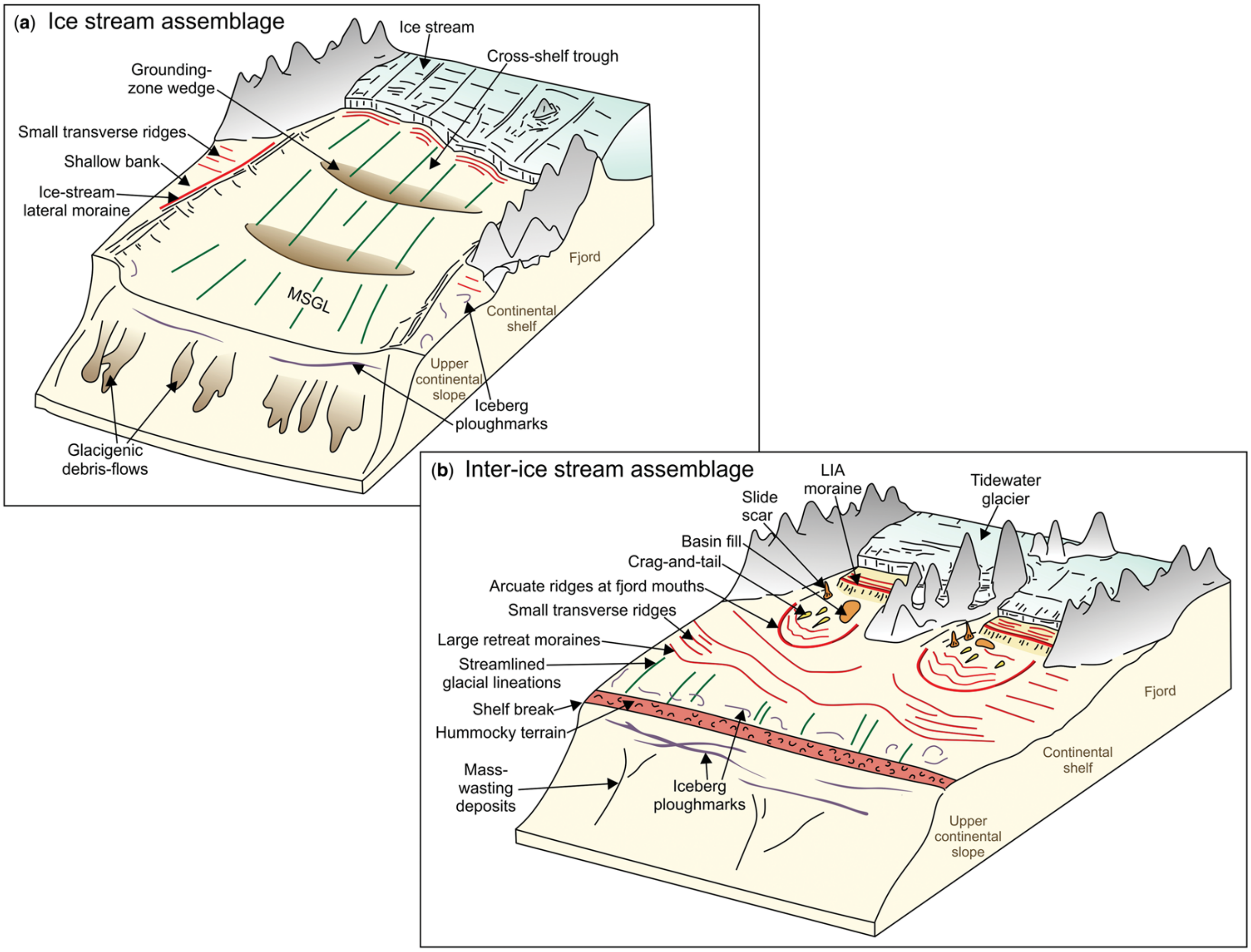


**Fig. 17.** The assemblages of submarine glacial landforms that characterize (**a**) ice stream and (**b**) inter-ice stream settings on high-latitude continental shelves (modified from Ottesen & Dowdeswell 2009). LIA, Little Ice Age; MSGL, mega-scale glacial lineations.

over very large areas in high-latitude cross-shelf troughs, where the acoustic interface with underlying stiff till provides an easily identified base to the deformable unit. For example, in Marguerite Trough, on the western Antarctic Peninsula, the thickness of deformation till beneath a former ice stream has been mapped over an area of 6000 km$^2$ and varies between about 4 and 19 m (Dowdeswell *et al.* 2004*a*; Ó Cofaigh *et al.* 2005*a*). In addition, the deformation till has been extensively sampled by coring to establish its mechanical properties, showing that the soft till is of low shear strength.

More generally, the spatial distribution of glacial landforms across the hundreds of kilometres of former ice-sheet beds that have been exposed by glacier retreat across polar shelves, but remain well preserved beneath hundreds of metres of water and sometimes draping sediment, has implications for the variable nature of the ice-bed interface. A consistent pattern of substrate and geomorphic change has been reported from many high-latitude shelves (Fig. 18) (e.g. Wellner *et al.* 2001, 2006; Canals *et al.* 2002, 2003; Anderson *et al.* 2014; Dowdeswell *et al.* 2014). Many inner-shelf areas have predominantly rock beds, sometimes separated by sedimentary basins, where glacial erosion has dominated and streamlined bedrock landforms such as roches moutonnées and whalebacks are present (e.g. Benn & Evans 2010; Dowdeswell *et al.* 2016*j*; Krabbendam *et al.* 2016). In addition, bedrock channels with undulating long profiles, formed by subglacial water flowing under high pressure, are common along with some evidence of wider basins that may have held larger subglacial water bodies (e.g. Domack *et al.* 2006, 2016; Nitsche *et al.* 2016*b*). On the inner shelf, there is often a down-flow transition in the nature of the substrate and its associated landforms where mixed-bed crag-and-tails and sedimentary drumlins predominate (Fig. 18). Further out on the shelf and extending to the shelf edge, the ice-sheet bed is exclusively sedimentary, related to the long-term deposition of glacial debris and the progradation of the continental shelf into deeper water. Here, the drumlins are typically replaced by highly elongate MSGLs, forming the upper surface of a deformation-till unit of metres in thickness (e.g. Anderson 1999; Dowdeswell *et al.* 2004*a*); meltwater channels are rarely observed, implying that they may be destroyed by continuous till deformation or that water is moving as part of the till itself. These gross changes in the nature of the ice-bed interface, along flowlines that extend across polar shelves hundreds of kilometres beyond the present locations of marine ice-sheet termini in the Arctic and Antarctic, provide an overview of the changing nature of the mechanisms that control the flow rate of both past and present ice sheets (Fig. 18). It is likely that ice sheets on full-glacial inner shelves were at the pressure-melting point in order to generate subglacial meltwater, but, especially where bedrock channels were present, efficient drainage restricted the rate of motion. The progressive shift to a sedimentary bed, the growing elongation

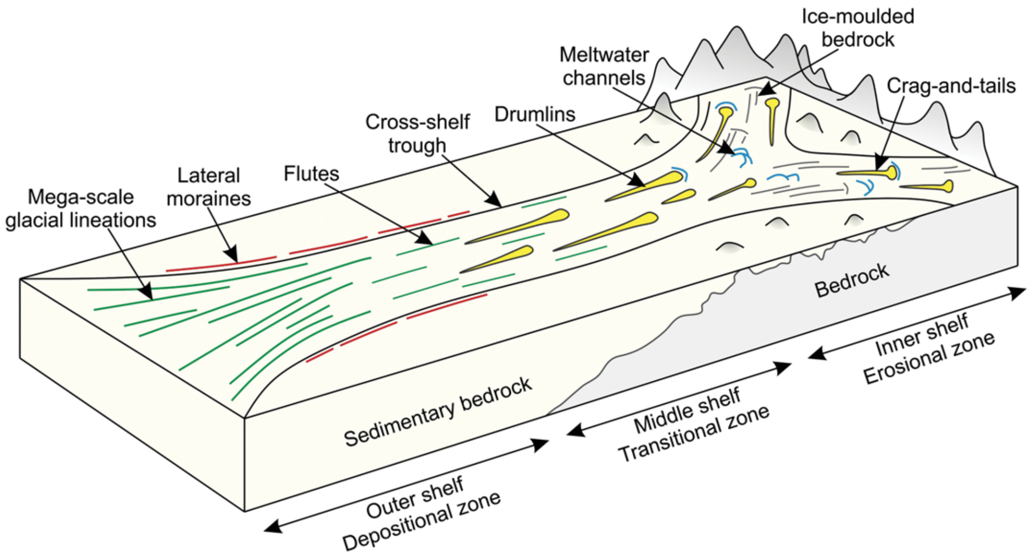


**Fig. 18.** Schematic diagram of the transition in submarine glacial landforms that takes place as ice flows from an interior ice-sheet drainage basin into a cross-shelf trough (modified from Wellner *et al.* 2001, 2006). Bedrock landforms are replaced by progressively more elongate streamlined sedimentary landforms as fast ice-stream flow takes place over a deformable bed.

ratios of streamlined sedimentary landforms and the possible lack of channels to drain the bed efficiently, all imply the speed-up of ice to form a fast-flowing ice stream (e.g. Wellner *et al.* 2001). By contrast, the lack of streamlined landforms on adjacent shallower banks between ice streams (e.g. Dowdeswell *et al.* 2016*h*; Klages *et al.* 2016), combined with the occasional presence of subglacially formed hill–hole pairs and transverse-to-flow retreat moraines, implies slower ice flow and, where hill–hole pairs are present, oscillations between a frozen and a thawed bed.

This information on the changing nature of the ice-bed interface, both along former ice streams and in zones of slower-flowing ice between them, provides an important basal boundary condition for numerical ice-sheet models; a comparable level of data from beneath modern ice sheets simply does not exist. By contrast, the observational datasets behind these general statements on the distribution of hard and soft beds on continental shelves are very comprehensive and based on systematic surveys and compilations of evidence from many high-latitude shelves in both the Arctic and Antarctic that have been affected by Quaternary ice sheets.

Finally, inferences on the likely past occurrence of floating ice shelves, although not their marginal extent beyond the grounding zone, can also be made from the distribution of various types of ice-marginal landform. Where transverse-to-flow moraine ridges of significant vertical expression (up to several tens of metres) exist, the implication is that vertical accommodation space at a grounded ice terminus is available to allow their development; that is, grounded tidewater glacier termini rather than floating ice shelves are present. By contrast, the often very subdued vertical topography associated with GZWs implies limited accommodation space beneath the confining ceilings of ice shelves extending beyond the grounding zone (Dowdeswell & Fugelli 2012; Batchelor & Dowdeswell 2015). This is important information for palaeoclimate reconstruction, in that comparisons between the modern distribution of Antarctic ice shelves and observed atmospheric temperatures show that ice shelves tend not to develop where mean annual temperatures exceed $-5^{\circ}$C (Robin 1979). The southward migration of the $-5^{\circ}$C isotherm down the western Antarctic Peninsula, where temperatures have risen by about 2.5°C in the past half century or so, has coincided with the break-up of fringing ice shelves (Vaughan & Doake 1996; Domack *et al.* 2005; Turner *et al.* 2005). Indeed, all the ice shelves on the Peninsula lying between the $-5$ and $-9^{\circ}$C isotherm have shown either significant recent retreat or total loss, noting that the $-9^{\circ}$C isotherm represents the approximate onset of summer surface-melting (Morris & Vaughan 2003).

## Discussion: meltwater beneath glaciers and ice sheets, and associated submarine landforms

### *A continuum of glacimarine environmental settings*

Glaciers and ice sheets reach the sea in a range of climatic and oceanographic settings today; these settings have also varied in their geographical distribution through successive Quaternary glacial–interglacial cycles (Fig. 19a, b). At one end of this continuum, relatively mild areas such as the fjords of Patagonia have water temperatures up to almost 10°C and mean annual temperatures consistently above zero, and large quantities of surface-derived meltwater are delivered to the margins of tidewater glaciers each year from snow- and ice-melt in a high-precipitation climate (e.g. Boyd *et al.* 2008; Dowdeswell & Vásquez 2013). By contrast, the coldest air and water temperatures are found offshore of the open-marine ice-sheet margin of East Antarctica, where water masses are consistently below 0°C, mean annual air temperatures are often $-15^{\circ}$C or lower, surface melting does not take place, and most mass loss is by iceberg production and the basal melting of floating ice shelves (e.g. Domack 1988; Rignot *et al.* 2013).

Given the very wide range of climates in which ice reaches the sea, it is no surprise that the influence of meltwater on rates of sediment supply, grain size and sorting, on the presence or otherwise of certain landforms, and on the nature and rate of development of submarine glacial landforms and sediments varies greatly (Fig. 19c). Sedimentation rates are generally highest where meltwater delivers sorted sediments to the glacimarine environment – rates reach tens of centimetres and even a metre or so per year close to the portals of surface-fed meltwater streams in SE Alaska (e.g. Powell & Molnia 1989; Powell 1990). By contrast, offshore of the frigid George V Land coast of East Antarctica, sediment delivery rates are very low, and sedimentation is predominantly by the slow rain-out of marine micro-organisms since even icebergs traversing the area melt only very slowly (e.g. Domack 1988).

### *Meltwater sources to the basal hydrological system*

Water beneath marine-terminating glaciers and ice sheets has several possible sources. First, ice-surface melting produces meltwater that penetrates to the bed via crevasses and moulins. The volume of surface meltwater delivered to marine ice margins will vary with air temperature, ice-mass hypsometry, the amount of snow available for melting each spring, and the amount of glacier ice

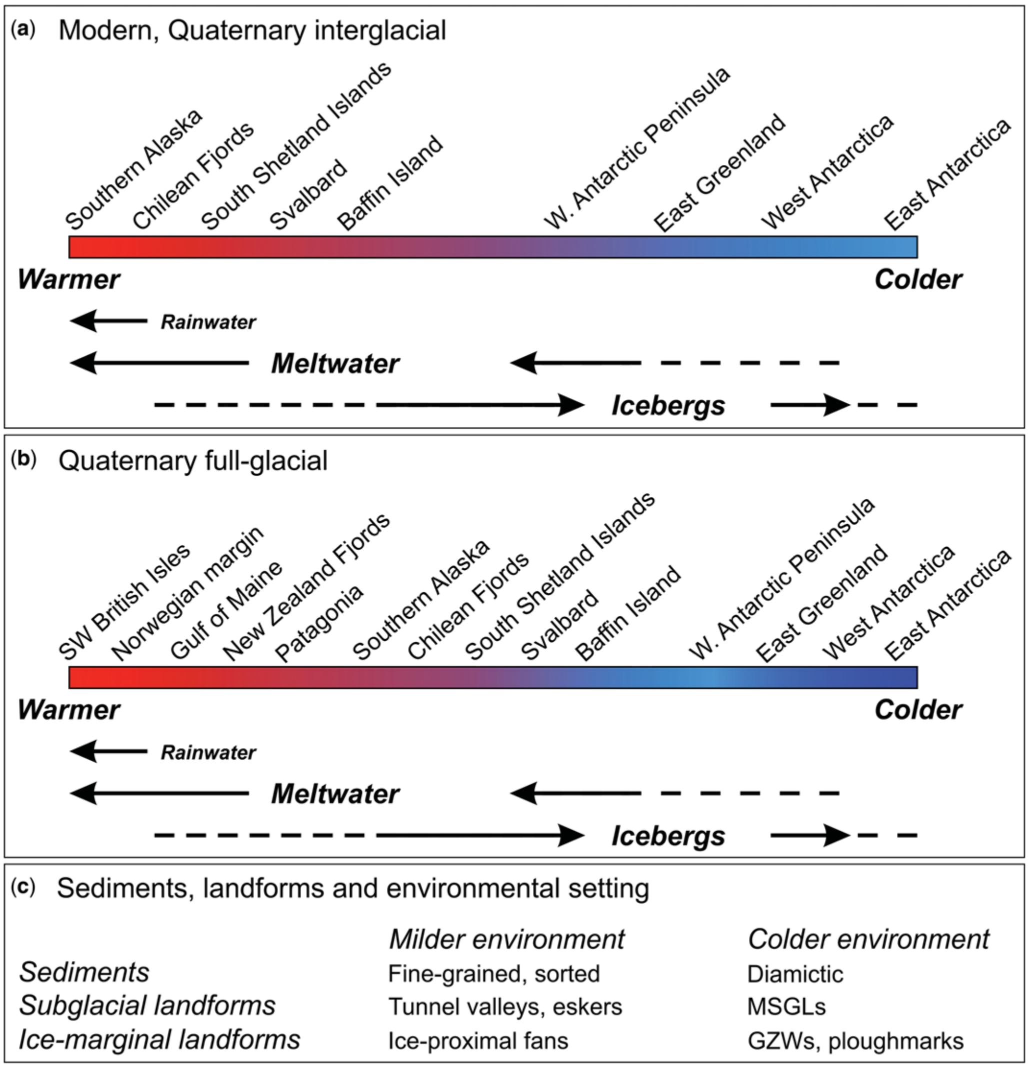


**Fig. 19.** The environmental continuum of glacimarine environments from the mildest to the coldest locations where ice reaches the sea. (**a**) Modern, Quaternary interglacial conditions. (**b**) Quaternary full-glacial conditions. (**c**) Some of the characteristic sediments and landforms associated with milder and colder glacimarine environments.

melted in the summer once any overlying snow has been removed. Surface melt rates can vary from just a few millimetres per year very close to the equilibrium line to ten or more metres per year close to the snouts of temperate glaciers. For the present-day ice sheets, measurements at *c.* 67°N on the western margin of the Greenland Ice Sheet show melt rates of up to 5 m $a^{-1}$ at *c.* 500 m elevation in the ablation zone, decreasing with elevation to the equilibrium line at around 1500 m (van de Wal *et al.* 2012); such values could be typical for the southern margins of mid-latitude ice sheets even at glacial maxima, and certainly during deglaciation. Secondly, water is produced at the bed by direct geothermal heating beneath relatively thick ice that brings the base to the pressure-melting point. Geothermal heat flux varies with crustal thickness and mantle activity and is, for example, relatively high beneath parts of the marine-based West Antarctic Ice Sheet where there is a subglacial volcanic province (Shapiro & Ritzwoller 2004). Estimates for the geothermal heat flux beneath Antarctica range from *c.* 50 mW $m^{-2}$ for the bulk of East Antarctica to *c.* 150 mW $m^{-2}$ for some regions in West Antarctica (e.g. Shapiro & Ritzwoller 2004). Given the latent heat of fusion of water, and assuming all of this heat is used to melt basal ice (whereas typically some heat will be lost into colder, overlying ice), these values yield estimates for geothermal melting of just *c.* 4–10 mm $a^{-1}$ – three orders of magnitude lower than possible surface melt rates. Some areas with higher geothermal heat flux, such as parts of the Ross Sea sector of Antarctica for example, may be more affected by basal meltwater than their latitudinal position alone might imply. Thirdly, basal meltwater is produced by strain heating which increases with the rate of ice flow. Meltwater derived from strain heating will be greatest where fast-flowing ice streams are present. Determining rates of frictional heating directly is very difficult, but numerical modelling studies suggest that values for frictional heating can range from a few tens of milliwatts per $m^2$ beneath slow-moving ice up to around *c.* 1–2 W $m^{-2}$ beneath fast-flowing ice streams (e.g. Golledge *et al.* 2014). These latter values would produce maximum rates of *c.* 90–180 mm of basal melt per year (again assuming all the available heat was used for melting).

### *Glacial landforms and varying meltwater availability*

A fundamental difference between ice masses is, therefore, whether or not surface meltwater contributes to water flow at the ice-sheet base. Where substantial surface melting takes place, this would normally be a much more significant contributor to the basal hydrological system than water from geothermal and strain heating. Given this, and that channel-based drainage is typically favoured by high subglacial discharges, it might be expected that subglacial channel networks would be best developed in relatively mild glacier-influenced marine settings. Similarly, ice-proximal fans, sourced at points where subglacial channels reach marine glacier termini, would also be more common where surface meltwater was available to augment more limited geothermal- and strain-produced basal sources. As well as the availability of water at the bed, it is also important to note that, irrespective of its source, subglacial water-flow paths are controlled by the subglacial hydraulic potential, itself a product of the basal and ice-surface topography, and subglacial water pressures. Water flowing beneath fast-flowing ice streams may be driven towards their lateral margins by the typical direction of basal water-pressure gradients (e.g. Dowdeswell *et al.* 2015). Channelized subglacial water may not, therefore, be expected along an ice-stream terminus, with possible exceptions being close to its lateral margins or when major outburst floods take place.

Submarine landforms indicative of channelized water flow beneath former ice sheets include tunnel valleys and smaller channels cut into subglacial sediments or bedrock (known as N or Nye channels), and the sinuous ridges – eskers – that represent the fill of channels that were incised upwards into basal glacier ice (known as R or Röthlisberger channels) (Fig. 4) (Röthlisberger 1972; Shreve 1972, 1985; Cuffey & Paterson 2010; Greenwood *et al.* 2016*a*). Meltwater channels appear to be relatively common in several submarine settings formerly occupied by ice sheets.

One setting where evidence of former subglacial channels is common is at or close to the southern margins of the Late Weichselian Eurasian Ice Sheet, in the North Sea at latitudes of about 50–60°N and in the Baltic Sea to the northeast (e.g. Huuse & Lykke-Andersen 2000; Stewart & Lonergan 2011; Greenwood *et al.* 2016*b*). In the Baltic, eskers are common depositional features, whereas a number of generations of larger and broader tunnel valleys are typically cut into North Sea sediments. At these latitudes, surface melting took place at the margins of the ice sheet even under full-glacial conditions, providing an abundant meltwater supply to the ice-sheet base which would have increased during deglacial warming. In much of the North and Baltic seas, especially beyond the Norwegian Channel and Skagerrak, ice retreat was also across a relatively shallow seafloor, which would have restrained mass loss by iceberg calving.

Tunnels valleys and eskers (Fig. 4), along with ice-proximal fans that are sometimes linked to so-called beads within esker systems, are indicators of relatively mild climatic conditions during Quaternary full-glacial periods. At these times, modelling suggests that the mid-latitude margins of ice sheets experienced abundant surface melting (e.g. Siegert & Dowdeswell 2002) that contributed to the formation of subglacial channels which filled with sorted and sometimes relatively coarse-grained sediments as discharge waned. Today, turbid meltwater plumes are observed regularly in summer beyond many tidewater glaciers in, for example, SE Alaska, Patagonia, Svalbard, Baffin Island, and parts of Greenland and the Antarctic Peninsula (e.g. Gilbert 1982; Powell 1990; Dowdeswell *et al.* 2015). The meltwater and suspended sediments making up these plumes, and providing sediments to build ice-proximal fans, are derived from the mouths of subglacial channels at the base of tidewater ice cliffs. Tidewater glaciers in these areas exhibit varying degrees of ice-surface melting that dominate the basal hydrological system and its development over a given summer (e.g. Dowdeswell *et al.* 2015).

Sediments deposited during the Late Ordovician glaciation of northern Africa about 440 million years ago also contain channel-like features that are similar in dimensions and sedimentology to the tunnel-valley systems of the Quaternary North Sea (Fig. 4b) (Le Heron 2016), and a number of ice-proximal fans have also been observed in outcrop (Hirst *et al.* 2002; Dowdeswell *et al.* 2015). This implies that the northern margin of the ice sheet that covered much of Africa during the Late Ordovician was probably located in a relatively mild climatic setting in which ice-surface melting took place, similar to the southern margins of the Eurasian Ice Sheet in the Late Weichselian. The sorted and sometimes relatively sandy tunnel-valley and fan sediments produced beneath these meltwater-rich ice-sheet margins are important hydrocarbon traps in both areas (e.g. Hirst 2012; Huuse *et al.* 2012).

A second, contrasting setting is in the inner-shelf areas of high-latitude ice-sheet/ice-stream systems that are hundreds of kilometres from the full-glacial ice margin at the continental shelf edge (Fig. 18). Here, systems of channels cut predominantly into bedrock are common in, for example, the deep waters of inner Pine Island and Marguerite bays and Getz Trough, Antarctica, at latitudes of 70–80°S (e.g. Lowe & Anderson 2003; Anderson & Fretwell 2008; Smith *et al.* 2009; Nitsche *et al.* 2013; Graham *et al.* 2016*a*, *b*). These bedrock channels have undulating long profiles, demonstrating that they were formed under pressure at ice-sheet beds; ice-sheet reconstructions suggest that LGM ice may have been up to several thousand metres thick at such locations. The source of this subglacial water is unlikely to be from the ice-sheet surface because, even today, the vast bulk of the Antarctic Ice Sheet experiences little or no surface melting even in summer – the climatic setting must have been even colder under full-glacial conditions. Even during deglaciation of many high-latitude continental shelves in Antarctica and Greenland, the bulk of mass loss probably occurred through the production of icebergs rather than by surface melting, as ice retreated inshore into deepening water that encouraged rapid calving. The meltwater eroding these bedrock channels probably came from more limited geothermal and strain heating, perhaps combined with the occasional drainage of subglacial lakes (e.g. Fricker *et al.* 2007). The presence of these bedrock channels on the inner parts of formerly ice-covered high-latitude shelves therefore holds little implication of mild conditions and surface meltwater availability, but rather it suggests basal meltwater production under thick full-glacial ice sheets where surface temperatures were probably considerably colder than those of today but ice was probably thicker and faster-moving. Some evidence of meltwater activity at or close to ice-sheet marine grounding zones has also been reported from GZWs on the continental shelves of Antarctica and West Greenland (McMullen *et al.* 2006, 2016; Dowdeswell & Fugelli 2012; Simkins *et al.* 2016).

By contrast with the characteristic presence of bedrock channels in many inner-shelf areas, cross-shelf troughs on the outer parts of many high-latitude continental shelves are dominated by elongate and streamlined sedimentary landforms, in particular MSGLs, and there is often no evidence of channels at all. This is typical of well-surveyed outer-shelf areas of Antarctica in, for example, Marguerite and Pine Island bays (e.g. Ó Cofaigh *et al.* 2002; Evans *et al.* 2006; Larter *et al.* 2009; Graham *et al.* 2010, 2016*a*, *b*; Livingstone *et al.* 2012, 2016). It is assumed that subglacial water, which is derived almost exclusively from geothermal and strain heating in full-glacial periods, is contained within the sediments themselves, contributing to the deformation of soft till under high pore-water pressure (e.g. Ó Cofaigh *et al.* 2005*a*).

Furthermore, in the outer parts of the Antarctic and northern Greenland shelves, and on the more southerly Norwegian shelf, the substrate is almost always made up of Quaternary sediments. Evidence of erosive channels or depositional channel-fill is limited on the Norwegian shelf north of the North Sea from about 62°N. Despite extensive coverage by seismic-reflection datasets, the pervasive networks of tunnel valleys observed on 2D profiles and in 3D seismic-reflection cubes in the Late Quaternary sediments of the North Sea are lacking (e.g. Rise *et al.* 2005; Ottesen *et al.* 2010). In addition, there is very limited evidence for channel development on the outer shelves off Greenland and Antarctica. It has been assumed that where basal meltwater is present beneath full-glacial and deglacial ice sheets on these high-latitude shelves, it is held largely within the sediment, contributing to subglacial deformation and motion when it is under pressure, and flows by Darcian processes. Water flow in small canals, or in a network where small channels are constantly evolving in association with soft sediments, below the resolution of most multibeam systems, may also be involved (Livingstone *et al.* 2016).

A possibly intermediate Late Quaternary environmental setting, between the relatively mild climate of the North Sea and the cold full-glacial shelves of Antarctica and northern Greenland, is that of the Barents Sea, where conditions during deglaciation from the LGM may have allowed significant surface meltwater to become available. At 75°N, in Kveithola Trough, grounding-line or ice-proximal fans, presumably deposited at the termini of subglacial streams, are present at the distal edges of GZWs (Bjarnadóttir *et al.* 2013; Bjarnadóttir & Andreassen 2016*b*). Deposition during Late Weichselian full-glacial and early deglacial times is inferred to have been by a combination of the delivery of deforming till along a line source at the grounding zone, and the intermittent presence of ice-proximal fans of sorted sediments that were point-sourced from the mouths of subglacial meltwater streams emerging at the grounding zone. Somewhat similar depositional features have been observed at several other locations on the Norwegian Sea and Barents Sea margins. Further east, several channels interpreted as tunnel valleys are identified on Sentralbanken (Bjarnadóttir *et al.* 2014). North of this location, at 77°N, a sediment apron has been mapped at a deglacial grounding line, although this feature may record either meltwater activity or debris flowage as deforming sediment was released at the grounding zone (Andreassen *et al.* 2014, 2016). In addition, Bjarnadóttir *et al.* (2014) point out that changes in meltwater abundance can lead to shifts in organization of the subglacial drainage system, with implications for changes in deposition at marine-terminating ice-sheet margins. On Murmanskbanken in the Barents Sea, for example, meltwater channels, suggesting relatively efficient meltwater drainage, breach some transverse retreat ridges. By contrast, a GZW on the bank did not show clearly identifiable channels, but water may instead have been escaping from a more distributed drainage system (Bjarnadóttir *et al.* 2014). Further south, on the Norwegian shelf at 65°N, there is a fan-like depocentre with an apparent point source that has been tentatively interpreted as formed by meltwater during deglacial ice retreat (Dowdeswell *et al.* 2016*i*). Such features are far from common at this latitude, however.

In addition, two further factors, which are less closely controlled by climate, may also have some influence on meltwater availability and grounding-zone sedimentation. First, wherever surface ice-dammed lakes or subglacial lakes drain to a marine ice margin as high-magnitude water discharge events (e.g. Fricker *et al.* 2007), these non-steady processes may produce significant depocentres of sorted sediment, often showing bedforms indicating high flow rates (e.g. Winsemann *et al.* 2009; Dowdeswell *et al.* 2016*i*). Secondly, where surface meltwater is unavailable, unusually high geothermal heat fluxes such as those beneath parts of the West Antarctic Ice Sheet affected by widespread Cenozoic rifting and volcanism (e.g. Blankenship *et al.* 1993; An *et al.* 2015), may provide an additional increment to the production of basal meltwater, yielding a possible explanation for the presence of the first soft-sediment-based channel to be reported on the Antarctic continental shelf in the western Ross Sea (Simkins *et al.* 2016).

### *Schematic model for marine glacier and ice-sheet termini: a continuum of meltwater influence*

In summary, it appears as though, following the early suggestion of Powell & Alley (1997), there may be an environmentally influenced continuum of grounding zones and their associated landforms and sediments where glaciers and ice sheets reach the sea (Fig. 20). At the mild end of the continuum, where surface meltwater is abundant and reaches the glacier bed, are large subglacial tunnel valleys, eskers and small ice-proximal fans composed of sorted sediments (Fig. 20a). The tunnel valleys and eskers are subglacial channel-fill and the fans are point-sourced from the mouths of well-developed subglacial hydrological systems fed largely by surface meltwater (e.g. Powell 1990; Dowdeswell *et al.* 2015). Between the channel mouths, transverse ridges are formed at tidewater glacier fronts (e.g. Dowdeswell *et al.* 2015), referred to by Powell & Alley (1997) as morainal banks. At the other, coldest, end of the environmental continuum are large diamictic GZWs which are formed at the grounding zone where ice shelves prevent vertical growth into ridges, and evidence of meltwater is either absent or limited to flow within till (Fig. 20d) (e.g. Dowdeswell & Fugelli 2012); here, the GZWs are formed largely of deforming till that may prograde as debris-flow deposits (e.g. Batchelor & Dowdeswell 2015).

On the continuum between these mild and cold end-members, the influence of surface meltwater diminishes progressively and, therefore, the smaller contributions from geothermal and strain heating become relatively more important (Fig. 20b, c). The examples of GZWs with small fans at their distal edges in the western Barents Sea are good illustrations of this intermediate position, where some surface meltwater may be available especially during climatic warming and associated regional deglaciation (Fig. 20b) (Bjarnadóttir *et al*. 2013; Bjarnadóttir & Andreassen 2016*b*). Where there is no surface water available, but geothermal and strain heating provide some basal water, GZWs are sometimes accompanied by limited signs of meltwater activity in the form of breach-points and, occasionally, small fans (e.g. McMullen *et al.* 2006). Finally, sedimentary fans produced by high-magnitude and low-frequency discharge events may be located at any latitude where subglacial or ice-dammed lakes are able to drain to a marine or lacustrine ice-sheet margin (Fig. 20e) (e.g. Piper *et al.* 2007; Winsemann *et al.* 2009).

## Further investigations of submarine glacial landforms

Going forward, there are a number of opportunities and some limitations to the study of submarine glacial landforms. First, progressively more of the high-latitude seafloor is being investigated to provide evidence on the planform and stratigraphy of increasingly remote parts of these often ice-infested seas. This allows us to continue to extend our understanding of the glacimarine sedimentary system and its variability in a range of environmental settings from the fjords of Patagonia and SE Alaska to the frigid waters offshore of East Antarctica.

Secondly, the advent of multibeam systems, allowing detailed mapping of seafloor geomorphology for the first time, represented a major, unprecedented breakthrough. The spatial resolution of these systems is becoming increasingly fine, and grid-cell sizes of surveys from surface vessels are increasingly at the scale of metres rather than tens of metres. This is a very significant trend, in that metre-scale imagery is important in refining our understanding of the detailed landforms and processes operating in glacimarine environments. The deployment of multibeam systems from remotely operated vehicles (ROVs) and autonomous underwater vehicles (AUVs) also allows geophysical equipment to be placed close to the seafloor, improving resolution and access to places which cannot otherwise be investigated, such as the cavities beneath ice shelves (Jakobsson *et al.* 2016*b*).

A third important opportunity is the increasing availability and quality of 3D seismic-reflection datasets from both modern and ancient glacier-influenced marine settings. The release of such datasets from the hydrocarbon-exploration and offshore-renewables industry, and the ability for high-resolution P-Cable methods to be utilized directly by academics, are both important. These datasets allow the variability of ice dynamics to be assessed through long periods of time, and the geometric development of high-latitude margins supplied cyclically with large quantities of glacial sediment over successive glacial–interglacial cycles to be investigated further. One limitation on this work is that, over the Quaternary, some palaeo-shelves have been removed or truncated by the erosive action of subsequent ice, although the occurrence of this is in itself useful information on the intensity of past glaciations.

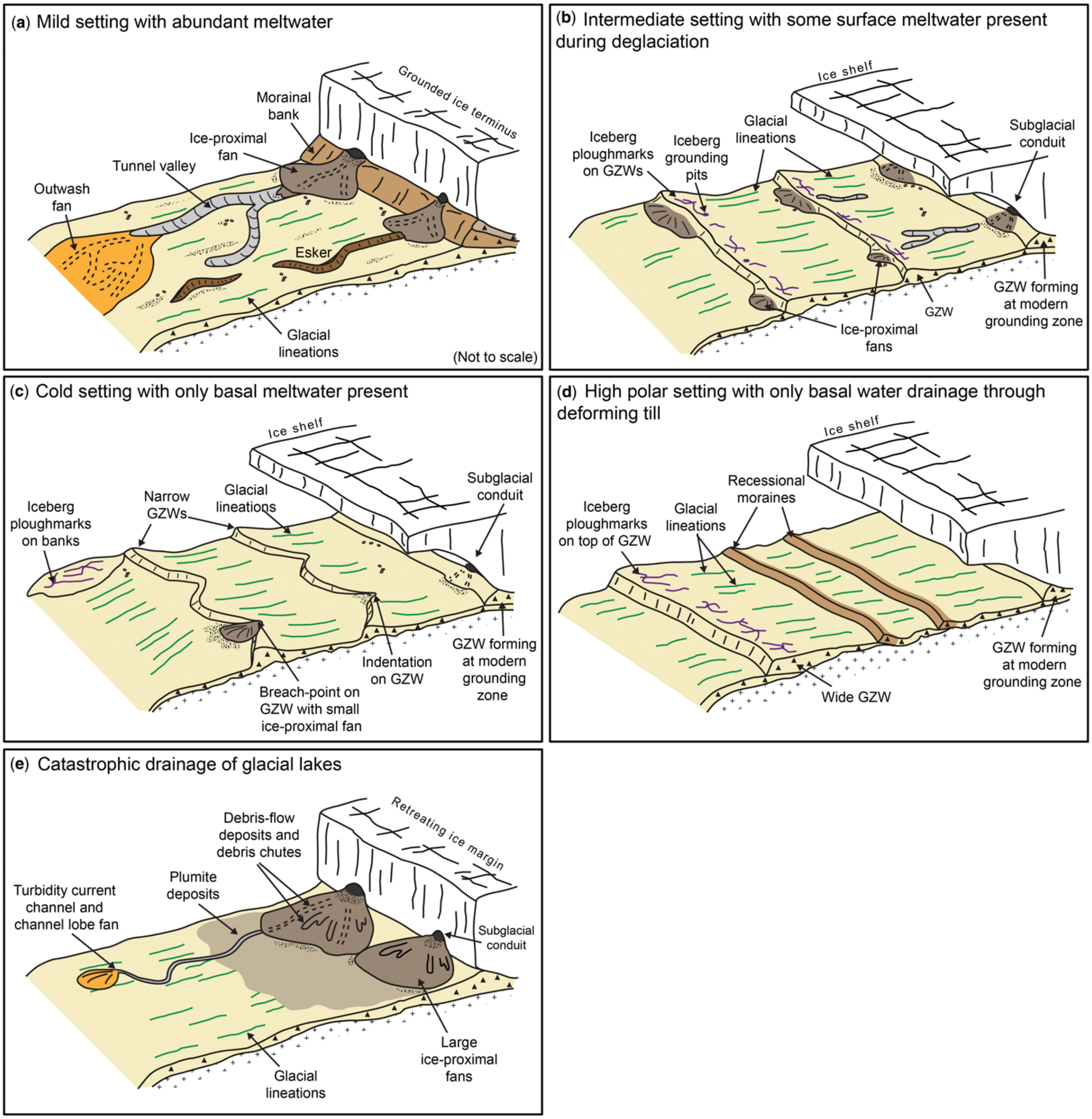


**Fig. 20.** Schematic diagram of the effects of changing surface-meltwater abundance on glacial landform development in marine ice-marginal settings. (**a**) Mild glacimarine setting with abundant surface meltwater. Ice-proximal fans, morainal banks and subglacially formed tunnel valleys and eskers are present (based in part on Powell 1990; Dowdeswell *et al.* 2015). Note the grounded tidewater ice cliff with no floating ice shelf beyond due to relatively mild ocean temperature. (**b**) An intermediate glacimarine setting, where surface meltwater is present during regional deglaciation, exemplified by Kveitola, western Barents Sea (modified from Bjarnadóttir *et al.* 2013). GZW, grounding-zone wedge. (**c**) GZW formed in a colder high-latitude glacimarine setting where only basal meltwater is available (based in part on McMullen *et al.* 2006; Dowdeswell & Fugelli 2012). (**d**) GZW in a high-polar glacimarine setting with no evidence of channels, where basal meltwater flows only in deforming till. (**e**) Large ice-proximal fan produced at the southern margins of a Quaternary ice sheet as a result of catastrophic drainage of an ice-dammed or subglacial lake (based in part of Winsemann *et al.* 2009). Note that the schematic diagrams are not drawn to the same scale.

Fourthly, there is a need for additional *in situ* observations of the nature and rate of contemporary processes involving landform development beneath ice sheets and on the adjacent seafloor. The use of high-resolution ice-penetrating radars and linked borehole investigations are examples relating to subglacial processes and the formation of streamlined sedimentary landforms (e.g. King *et al.* 2007, 2009). The role of cascading dense-water and turbidity currents in settings ranging from delta fronts to the continental slope is another area where continuing process studies are important (e.g. Hughes Clark 2016).

Fifthly, there is a need for greater chronological control on the submarine sediments and landforms produced by former glaciers and ice sheets. This is important in order to improve our knowledge of rates of change and build-up in several areas, for example: retreat rates of marine ice margins during deglaciation and the consequent implications for sea-level change; the timing of switches in ice-stream direction in response to changing flow dynamics and external environmental drivers including air and ocean temperature and precipitation; and the rates of debris delivery and build-up of sedimentary depocentres and individual landforms. The establishment of key sites for absolute dating, driven by the distribution of submarine glacial landforms and related stratigraphies, is an important approach (e.g. Ó Cofaigh *et al.* 2014), and is necessary if sense is to be made of the increasingly complex patterns of ice-sheet change implied by glacial landform suites in, for example, the Barents Sea, the western Ross Sea, around the British Isles, and on a number of other high-latitude margins (e.g. Andreassen *et al.* 2008, 2014; Bjarnadóttir *et al.* 2013; Clark *et al.* 2012; Greenwood *et al.* 2012; Anderson *et al.* 2014). The shifting pattern of ice flow implied by several generations of cross-cutting MSGLs in McMurdo Sound, western Ross Sea, provides a clear example (Fig. 16).

Finally, the links between ancient, Quaternary and modern glacimarine environments, landforms and sediments provide ongoing opportunities to improve descriptions and increase understanding. Sediment cores and acoustic stratigraphies can be obtained offshore of modern tidewater glaciers and marine-terminating ice sheets, but detailed direct sedimentological data are limited. By contrast, ancient systems such as the glacial rocks of the Late Ordovician in North Africa or the Permian in Antarctica provide major outcrops with good exposure where very detailed sedimentological investigations can take place. The same is, of course, the case for uplifted late Quaternary glacimarine sediments and landforms formed in relatively shallow water that have become emergent above sea-level (e.g. Domack 1984; Ó Cofaigh 1999; Lindén & Möller 2005), where the environmental and topographic context is much simpler to unravel than in ancient glacial rocks. The iteration between older outcrops and modern glacimarine settings where processes can be observed directly is an important area of continuing research. The understanding of modern process environments and their outcomes in terms of landforms, sediments and broader-scale architecture also remains important in the search for bodies of high-porosity sorted sediments in ancient glacial systems, given their potential as hydrocarbon reservoirs.

## Summary

The variety of submarine glacial landforms exposed at the seafloor or imaged on buried continental shelves, illustrated here (Figs 2–9; Table 1) and throughout the *Atlas of Submarine Glacial Landforms*, provides compelling evidence not only for the extent, growth and decay of past ice sheets through the Quaternary and in more ancient glacial systems, but also for the complexities and changing dynamics that are involved (Fig. 16). Although often well preserved in the Quaternary record, these glacial landforms have been reworked in some areas by the action of iceberg-keel ploughing (Fig. 9) and by other marine-geological processes including a variety of slope-failure types and currents (Figs 10 & 11). In addition, although much of the global seafloor is relatively little affected by humankind, landforms in some high-latitude fjords and shelves have been modified by, for example, trawling, waste disposal and hydrocarbon-related activities (Fig. 12).

Landforms produced by the action of ice are formed subglacially, ice-marginally and by icebergs and sea-ice drifting in the adjacent ocean waters. These landforms range in volume from huge TMFs, which are some of the largest morphological features seen on continental margins anywhere on Earth, to berms often less than a metre high produced as iceberg keels plough through soft shelf sediments (Fig. 13). The time taken to form the very wide variety and dimensions of glacial landforms ranges from hundreds of thousands to millions of years for TMFs to only minutes or hours for berms at the edges of individual iceberg ploughmarks (Fig. 13). We do, however, know more about the geometry and volume of glacier-related landforms than we do about the time they take to form, due in part to the difficulties of dating glacigenic sediments that are often largely devoid of organic material – although Optically Stimulated Luminescence methods may be an alternative to radiocarbon in some cases (e.g. Mellett *et al.* 2012).

Glacial landforms, and assemblages of landforms, preserved on the seafloor and buried in Quaternary depocentres, are clear indicators of the extent and flow directions within former ice sheets (Fig. 14); streamlined glacial landforms are especially useful in reconstructing past flow directions (Fig. 2) and transverse-to-flow ridges of varying dimensions often mark maximum ice extent and still stands during deglaciation (Fig. 6). Very large glacial landforms, such as fjords, cross-shelf troughs and TMFs, which often build up over a number of glacial–interglacial cycles, indicate the sustained action of glaciers and ice sheets through much of the Quaternary (Fig. 15). Examination of the glacimarine sedimentary record at increasingly high resolution has also revealed the complexity of ice-sheet dynamics during retreat from full-glacial ice at the shelf edge. Cross-cutting relationships between glacial landforms (Fig. 16), often produced over just a few thousand years or less, imply shifting interactions and feedbacks between sea-level rise, warming climate and the changing relationship between mass loss by iceberg production and meltwater runoff as deglaciation proceeds across shelves and fjords of varying depths.

The assemblages of glacier-influenced landforms, and their distribution across whole fjord–shelf–slope systems, also contain significant information on past ice flow and the climatic setting in which former ice sheets grew and decayed. There is a distinctive difference between the predominantly streamlined subglacial landforms that indicate the presence of past ice streams in cross-shelf troughs (Fig. 17a) and the mainly transverse-to-flow landforms that mark the slower retreat of ice across shallower banks between ice streams (Fig. 17b). The transition from relatively slow-moving ice in the interior drainage basins of ice sheets, to fast-flow in cross-shelf troughs, is also indicated in the landform record; there is a shift from channels, crag-and-tails and other streamlined bedrock landforms on the crystalline rocks that often make up the coastal zone of high-latitude margins, to almost exclusively streamlined sedimentary landforms of high elongation ratio on the outer shelf (Fig. 18).

Finally, the glacimarine system encompasses modern environments that range from the relatively mild climatic and oceanographic conditions of Patagonia and SE Alaska, to the very cold air and water temperatures of East Antarctica (Fig. 19). The glacial and glacimarine processes that operate across this wide range of environmental settings are dominated by surface-meltwater runoff and the deposition of predominantly sorted sediments in relatively milder lower latitudes where ice reaches the sea, and by iceberg calving and more restricted basal meltwater production at higher latitudes. The contrasts in both landforms and sediments across this range of present and past glacimarine settings can be seen not only in modern and Quaternary examples (Fig. 20), but also in the meltwater-dominated conditions indicated by channelized landforms and sorted sediments at the margins of the ancient ice sheet that covered northern Africa during the Late Ordovician.

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